ECONOMIC GEOLOGY

VOL. XXXVII SEPTEMBER–OCTOBER, 1942 No. 6

SOME CONCEPTS ON THE GEOLOGY OF QUICKSILVER DEPOSITS IN THE UNITED STATES.[1]

CLYDE P. ROSS.

CONTENTS.

Abstract 439
Foreword 440
Distribution 443
Kinds of lodes 445
Mineralogy 450
Metallic minerals 450
Gangue minerals 451
Structure 454
The mineralizing solutions 459
Relations to water of meteoric origin 460
Conclusion 464

ABSTRACT.

The concepts as to the origin of quicksilver deposits in the United States here outlined are based largely on data obtained by members of the Federal Geological Survey during its Strategic Mineral Investigation, still in progress, and similar earlier studies. Consideration of the distribution of such deposits throughout the country leads to the conclusion that they are rarely associated with deposits of other metals, and that each of the principal districts is dominated by a single outstanding mine, seven of which together have yielded fully 80 per cent of the nation's total production. The principal kinds of deposits include (1) the roughly tabular lodes in California, associated with bodies of altered serpentine, (2) those in Texas, most of which are in fractures opened by tension and enlarged by ground water solution in limestone beneath a capping clay unit, (3) those in western and central Oregon mostly in ill-defined fractures and irregular faults in altered Tertiary sedimentary and volcanic rocks, (4) the "opalite"

[1] Published with the permission of the Director, Geological Survey, U. S. Dept. of the Interior, Washington, D. C.

deposits in southeastern Oregon, northern Nevada, and other localities, (5) lodes in Arkansas, mainly in sandstone and associated with tear faults and cross folds above an overthrust. In addition there are several other less important varieties. Most of the lodes of all varieties are aggregates of fillings of numerous spaces that were open or under slight pressure at the time of mineralization. Most cinnabar coats fractures or other pre-existing openings, many individually minute, and the mineral has rarely replaced the wall rocks extensively. In many deposits the cinnabar represents a late stage in the history of mineralization and the wall rocks have been extensively altered either earlier in the formation of the lodes or during some previous hydrothermal episode. In places, the cinnabar is concentrated in structurally favorable situations of diverse kinds. Mineralization is effected by rising solutions of alkaline sulphides that probably have been much modified by the loss of different constituents at greater depths. In some places they have mingled with ground water so that the gangue minerals deposited from them vary with the character of the nearby rocks. Precipitation is caused by such factors as increase in acidity and drop in temperature and may be aided by encountering surface water percolating downward.

FOREWORD.

THE geology of the quicksilver deposits in the United States was studied by Becker and summarized in some detail in a report published in 1888.[2] At that time nearly all of the known valuable deposits were in California. For some time thereafter, comparatively slight advances were made in concepts pertaining to the geology of this liquid metal, although deposits were opened in numerous new localities, some of which had geologic features distinctly different in detail from those described by Becker. In a paper first published in 1930, Schuette[3] presented his ideas on the subject, emphasizing particularly the role played by structural traps in the localization of ore shoots. From approximately that time to the present a number of reports on outstanding districts and groups of districts have appeared. The states of Arizona,[4] Cali-

[2] Becker, G. F.: Geology of the quicksilver deposits of the Pacific slope, with an Atlas. U. S. Geol. Surv. Mon. 13, 486 pages, 1888.

[3] Schuette, C. N.: Occurrence of quicksilver ore bodies. A. I. M. E. Tech. Pub. 335, 88 pp., 1930, Transactions, 1931: 403–488, 1931.

[4] Lausen, Carl, and Gardner, E. D.: Quicksilver resources of Arizona. Ariz. Bur. Mines, Univ. Ariz., Bull. 122, 112 pages, 1927.

fornia,[5] Oregon,[6] and Texas,[7] have issued summary descriptions of their quicksilver deposits which include much geologic information. The California bulletin cited is based in part on two similar, earlier bulletins.[8] The investigations by the Federal Geological Survey under the Strategic Minerals program begun in 1939 and those under allotments from the Public Works Administration in 1934 and 1938 include work in a large number of quicksilver mining districts. The present writer has been the author of several of the reports listed below and has visited each of the other districts described, mostly in company with the geologists who described them. Several other reports on quicksilver districts are now in course of preparation in the Geological Survey. In addition to being familiar with most of these reports, the writer has, at one time or another, visited nearly every quicksilver mining district in the United States.

LIST OF REPORTS RESULTING FROM RECENT STUDIES BY THE GEOLOGICAL SURVEY.

Reed, J. C., and Wells, F. G.: Geology and ore deposits of the southwestern Arkansas quicksilver district. U. S. Geol. Surv. Bull. 886–C: 15–90, 1938.

Dreyer, R. M.: Goldbanks mining district, Pershing County, Nevada. Univ. Nev. Bull. 34: No. 1, 38 pp., Jan. 15, 1940.

Roberts, R. J.: Quicksilver deposits of the Bottle Creek district, Humboldt County, Nev., a preliminary report. U. S. Geol. Surv. Bull. 922–A: 1–30, 1940.

[5] Ransome, A. L., and Kellogg, J. L.: Quicksilver resources of California. Cal. Jour. Mines and Geol., Cal. Div. of Mines, 35: 353–486, Oct. 1939.

[6] Schuette, C. N.: Quicksilver in Oregon. Oregon Dept. Geol. and Min. Indus., Bull. 4, 172 pages, 1938.

[7] Baker, C. L.: Mercury minerals and ores. Geol. of Texas, Univ. Texas Bull. 3401: 511–558, 1935. Ross, C. P.: Preliminary report on the Terlingua quicksilver district, Brewster County, Texas. Geol. of Texas, Univ. Texas Bull. 3401: 558–573, 1935.

[8] Forstner, Wm.: Quicksilver resources of California. Cal. State Min. Bur., Bull. 27, 273 pages, 1903. Bradley, W. W.: Quicksilver resources of California. Cal. State Min. Bur. Bull. 78, 389 pages, 1918.

Ross, C. P.: Quicksilver deposits of the Mount Diablo district, Contra Costa County, Calif. U. S. Geol. Surv. Bull. 922–B: 31–54, 1940.

Roberts, R. J.: Quicksilver deposit at Buckskin Peak, National mining district, Humboldt County, Nev. U. S. Geol. Surv. Bull. 922–E: 115–133, 1940.

Ross, C. P.: Quicksilver deposits of the Mayacmas and Sulphur Bank districts, Calif. U. S. Geol. Surv. Bull. 922–L: 327–353, 1940.

Eckel, E. B., Yates, R. G., and Granger, A. E.: Quicksilver deposits of San Luis Obispo County and southwestern Monterey County, Calif. U. S. Geol. Surv. Bull. 922–R: 515–580, 1941.

Ross, C. P.: Some quicksilver prospects in adjacent parts of Nevada, California, and Oregon. U. S. Geol. Surv. Bull. 931–B: 23–37, 1941.

Ross, C. P.: The quicksilver deposits of the Terlingua region, Texas. Econ. Geol., 36: 115–142, 1941.

Ross, C. P.: Quicksilver deposits in the Steens and Pueblo Mountains, southeastern Oregon. U. S. Geol. Surv. Bull. 931–J: 227–258, 1942.

Dane, C. H., and Ross, C. P.: The Wild Horse quicksilver district, Lander County, Nevada. U. S. Geol. Surv. Bull. 931–K, in press.

Yates, R. G.: Quicksilver deposits of the Opalite district, Malheur County, Oregon, and Humboldt County, Nevada. U. S. Geol. Surv. Bull. 931–N, in press.

Ross, C. P., and Yates, R. G.: The Coso quicksilver district, Inyo County, California. U. S. Geol. Surv. Bull. 931, in preparation.

Dane, C. H., and Ross, C. P.: Quicksilver deposits near Weiser, Washington County, Idaho, in preparation.

The present summary is based on the experience above outlined, and on discussion of the problems involved with so many geologists and operators, that individual acknowledgment of the aid received is impracticable here. The paper should be regarded

merely as a report of progress in the growth of concepts that are expected to be modified as investigations proceed. The generalizations here expressed are intended to apply only to deposits within the United States. Many of the observations on which this paper is based are summarized in the reports referred to above but as most of these were prepared mainly for the purpose of outlining the economic possibilities of the different districts, abstract concepts and many of the facts relative thereto have been omitted from them.

As the present paper deals solely with geologic matters, the term quicksilver is employed throughout in accordance with usage among mining men. Once the metal has left the mine it is almost invariably called mercury. The mineralogists generally follow chemists in speaking of " mercury" rather than " quicksilver."

DISTRIBUTION.

Quicksilver deposits are widely distributed in the United States as is evident from Fig. 1. Among the significant features of the distribution is the small number of outstanding producers and the tendency for quicksilver mining districts to be remote from those of other metals. Certain kinds of deposits are associated with particular kinds of rocks, primarily because such rocks for diverse reasons provide especially abundant openings. In the Coast Ranges in California, for example, many deposits are in brecciated serpentine, rendered brittle by alteration, but some are in other kinds of fractured or brecciated rock. Similarly the distribution of quicksilver deposits in other areas depends more on the extent to which open spaces were accessible to the solutions than on the particular kinds of country rocks available.

The principal productive localities are in western California, southwestern and central Oregon, and the Big Bend of the Rio Grande in west Texas, but southeastern Oregon, Arkansas, numerous districts in Nevada and Idaho, and some in Arizona, Washington, and other states also contain mines, most of which have come into prominence rather recently. Nearly all the larger

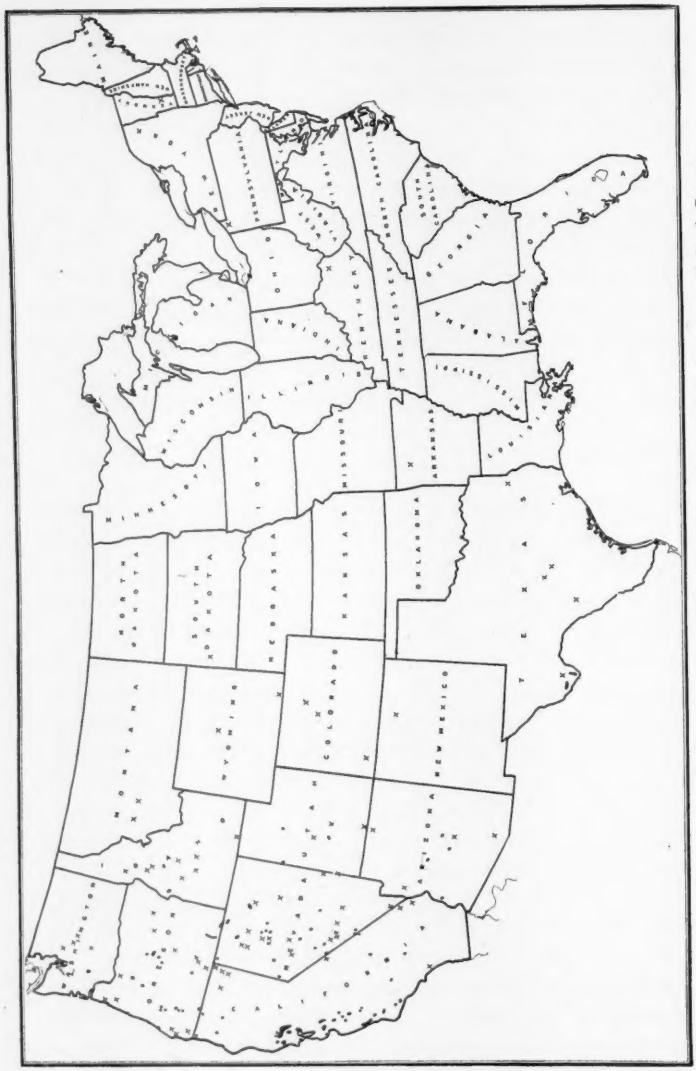


FIG. 1. Map showing the distribution of quicksilver deposits in the United States.

producers have been known for 40 years, and the principal ones for 80 years or more. There are several hundred productive mines, but fully 80 per cent of the nation's total yield has come from the New Almaden, New Idria, Oat Hill, Knoxville, Sulphur Bank, and Great Western in California, and the Chisos in Texas. The first two listed, both productive for more than 85 years, appear to have yielded fully half of the total. In nearly all of the more productive districts the production from a single mine overshadows that of the rest of the district. No district that has been prominent in the quicksilver industry has had any large production of other metals and few are close to mining districts of other kinds. In a word, quicksilver is a "lone wolf" among metals.

KINDS OF LODES.

The quicksilver deposits of the United States differ among themselves in multifarious details but most of them can be grouped in a few categories, primarily on the basis of the kinds of associated rocks, the fracturing of which has influenced the form and mineral content of the lodes. The outline below is intended as an aid in the descriptions that follow, but it is not so complete as to embrace all varieties of deposits in this country. Available data are not sufficient to make possible such a comprehensive classification, even if it could be compressed into the space available here. The principal deposits include those associated with altered serpentine, mainly in California; those localized beneath impermeable beds at a particular stratigraphic horizon, best exemplified in Texas; those scattered through zones in which sedimentary and locally volcanic rock had previously been impregnated with carbonates and quartz, mainly in western Oregon; those of the "opalite" type; those related to cross folds associated with thrusts as in Arkansas.

The best known deposits, those that up to the present have yielded most of the country's supply of quicksilver, are in the Coast Ranges in California. Similar lodes are reported in the extreme southwest part of Oregon. Most are in serpentine or,

FIG. 1. Map showing the distribution

more commonly in its alteration product, silica-carbonate rock, which is also called calc-silica rock and quicksilver rock. Some are in the sandstone and chert of the Jurassic (?) Franciscan formation, commonly close to intrusive masses of serpentine. Many of the intrusive masses are roughly sill-like but now steeply inclined, and ore shoots tend to lie close to their under sides. Exceptionally the lodes are in Cretaceous and Tertiary rocks and a few extend into Recent basalt. These exceptional lodes include some of the most productive orebodies, for example, at Sulphur Bank and New Idria. Most of the deposits in California are rather steeply inclined tabular or lenticular bodies in which altered country rock is the principal component.

Silica-carbonate rock, as that term is used in the present paper, is an alteration product of serpentine. It varies in many details and merges so intimately with the serpentine that it has rarely been distinguished on maps. Some varieties have been described in detail by Knopf.[9] Chalcedony, opal, and magnesian carbonates in widely varying proportions are the principal components, although some serpentine generally remains. Minor constituents such as iron and manganese oxides and nickel silicates impart different colors to the rock. Much of the rock, especially in and near quicksilver ore shoots, is intensely brecciated. Silica-carbonate rock is widespread in the serpentine masses of the Californian Coast Ranges, especially north of San Francisco, and is by no means confined to the immediate vicinity of quicksilver ore-shoots. Alteration products of this general sort are present in serpentine masses in many regions,[10] and many of them are independent of ore deposits of any kind. Somewhat similarly altered serpentine, regarded as related to the Jurassic (?) gold quartz veins, is reported to be plentiful in parts of the Sierra Nevada.[11] This is evidence that such alteration did take place, in

[9] Knopf, Adolph. An alteration product of Coast Range serpentine. Univ. of Cal. Pub., Dept. of Geol. Bull. 4: 425–430, 1906.

[10] Benson, W. N.: The origin of serpentine, a historical and comparative study. Amer. Jour. Sci., 4th Ser., 46: 710–714, 1918.

[11] Knopf, Adolph: The mother lode system of California. U. S. Geol. Surv. Prof. Pap. 157: 20, 44, 1929. Ferguson, H. G., and Gannett, B. W.: Gold quarts veins

that region at least, soon after serpentinization, although the gold quartz veins of the Sierra Nevada have no counterpart in the Coast Ranges.

The writer's interpretation is that the silica-carbonate rock of the Coast Ranges was formed during a late stage in the series of hydrothermal processes in which the original ultra-basic rock was converted into serpentine.[12] It has evidently been a favorable host rock for quicksilver lodes largely because it is brittle. In ore zones the silica-carbonate rock has been modified by later hydrothermal processes that were related to cinnabar deposition, and not to serpentinization. In places it is difficult to distinguish between the products of the different changes. The "twice-silicified reefs" of the Mayacmas and other districts in California are regarded as among the results of modification of silica-carbonate rock during quicksilver deposition.

The quicksilver lodes in the Coast Ranges of California, with a few notable exceptions, fill groups of open fractures and breccias. Among the more prominent Californian ores that are not in fracture or breccia fillings are the ore in the quarries of the Sulphur Bank mine, Lake County, most of which is in a part of a Recent lava flow made extremely permeable by hydrothermal alteration, and the part of the ore in the Oceanic Mine, San Luis Obispo County, that is disseminated in Miocene sandstone.

In the Big Bend area of the Rio Grande, which has placed Texas second in total aggregate production among quicksilver mining states,[13] many of the lodes are near the base of the impervious Del Rio clay (Lower Cretaceous); some are at higher horizons, and a few are in intrusive masses of early Tertiary age, thought to be genetically related to the mineralization. The solutions that formed these deposits penetrated along fractures

of the Allegany district, California. U. S. Geol. Surv. Prof. Pap. 172: 14, 1932. Johnson, W. D., Jr.: The gold quartz veins of Grass Valley, California. U. S. Geol. Surv. Prof. Pap. 194: 12, 1940.

[12] Ross, C. P.: Quicksilver deposits of the Mayacmas and Sulphur Bank districts, California. U. S. Geol. Surv. Bull. 922–L: 332–333, 1940.

[13] Ross, C. P.: The quicksilver deposits of the Terlingua region, Texas. ECON. GEOL., 36: 115–142, 1941.

opened by tension and, in the limestone beneath the Del Rio, enlarged by ground water solution. Many of these deposits are along fractures of northwest trend, which belong to a regional joint system, but some, especially at horizons above the Del Rio, are related to faults not a part of this system. The numerous factors that have affected deposition in this region have resulted in diversity in the form of the deposits.

In the principal quicksilver districts of western Oregon,[14] the cinnabar is distributed along ill-defined fractures within parts of the dominantly sedimentary beds of early Tertiary age, which at an early stage of mineralization were impregnated with calcite, ankerite, siderite, and fine-grained quartz. Some of the rock thus altered was originally lava. In mineral composition these altered rocks resemble the silica-carbonate rock of California, but for the most part they retain the appearance of the original rock and apparently none of them have the opaline luster characteristic of much of the silica-carbonate rock derived from serpentine. The introduced silica in the rock in Oregon is confined to the immediate vicinity of a network of fractures in softened rock and therefore results in a distinctive pattern of narrow, raised ribs, generally iron-stained, on weathered surfaces. The cinnabar in some lodes is disseminated through the rocks; in others it lines minor fractures.

The variety of quicksilver lode that has come to be known as the "opalite" type differs in several respects from those described above.[15] The deposits in southeastern Oregon and northern Nevada are the best known representatives of this type, but others are scattered throughout Nevada, southwestern Idaho, and parts of California. Typically they consist of rudely tabular and roughly horizontal bodies of fine-grained silica that were formerly thought to have been deposited at the surface by hot spring waters. Although certain opalite bodies had such an origin,

[14] Wells, F. G., and Waters, A. C.: Quicksilver deposits of southwestern Oregon. U. S. Geol. Surv. Bull. 850, 56 pages, 1934.

[15] Yates, R. G., and Roberts, R. J.: The "opalite" type of quicksilver deposit (Abstract). ECON. GEOL., 36: 839, 1941.

most of them are now known to result from replacement, largely of Miocene clastic beds and lavas, by silica which may have been deposited mainly as opal but is now largely chalcedony with subordinate quantities of other forms of silica. In a few deposits opal is still the principal form of silica present. Most of the cinnabar, accompanied by a little silica, occupies fractures and other openings, many of them minute, in the silicified rock. In many of the deposits clay minerals are present, and in some, such as those near Weiser, Idaho, they are plentiful. It is evident from the above summary that opal is not now a major constituent of many of the lodes commonly called opalite deposits; however, the term is in common use. Its use is supported by the fact that opal is or once was present in many of the deposits and that nearly all the silica in them is fine-grained. In some places steep feeder channels have been found. These contain material similar to that of the opalite deposits above, but most of them contain little ore.

In Arkansas [16] most of the cinnabar deposits are in a thrust block that is cut by tear faults and deformed by cross folds. Many of the ore shoots are related to the cross folds in certain fractured sandstone beds. The cinnabar is accompanied by quartz and the clay mineral dickite. Reed and Wells are inclined to regard the lodes as of middle Pennsylvanian age, but they might, perhaps, be younger.

The United States contains several other kinds of quicksilver deposits. Those in central Oregon are in lower Tertiary tuff and lava, without the introduced silica and carbonate characteristic of the wall rocks of deposits in the western part of the state. In eastern Harney County, Oregon, there are long siliceous reefs containing quicksilver-bearing tetrahedrite and cinnabar that as yet have been almost non-productive. In central and southern Nevada the deposits differ in details in accord with the character of the enclosing rocks, which include Tertiary volcanic rocks and Mesozoic and Paleozoic limestone and schist. The deposits in

[16] Reed, J. C., and Wells, F. G.: Geology and ore deposits of the southwestern Arkansas quicksilver district. U. S. Geol. Surv. Bull. 886–C: 15–88, 1938.

Arizona are mainly in shear zones in pre-Cambrian schist, and much of the cinnabar coats schistose partings.

MINERALOGY.

Metallic Minerals. The mineralogy of most quicksilver lodes is simple. Perhaps 95 per cent of all the quicksilver mined in the United States has come from cinnabar, and cinnabar is also present in most of the comparatively few ore shoots in which native quicksilver, metacinnabar, or the oxychloride minerals are plentiful. Cinnabar is generally one of the latest minerals to crystallize. Much of the cinnabar that is clearly of hypogene origin, with the exception of part of that in the opalitic deposits, is sufficiently coarse grained for crystal faces to be visible. The variety commonly called "paint" is finer-grained and more earthy. It lines fractures in orebodies and their wall rocks. Although some paint cinnabar is undoubtedly hypogene, some may be supergene. Both the more coarsely crystalline and paint varieties darken superficially on exposure, sufficiently to hinder prospecting by concealing the characteristic vermillion color of the mineral. As Dreyer [17] has noted, the cause of this discoloration is unknown. Quicksilver-bearing tetrahedrite has been reported from several localities, mostly in Oregon and Nevada, but in most of these the quicksilver content of the lodes is small. Some of the deposits that contain this mineral, particularly in Harney County, Oregon, may eventually prove to be valuable sources of quicksilver. Most deposits of quicksilver-bearing tetrahedrite also contain paint cinnabar, at least in part derived from the tetrahedrite. Quicksilver tellurides are reported from several localities in Colorado, and selenides from Utah and a few places in California, but apparently no quicksilver has yet been produced commercially from tellurides and little from selenides.

In many cinnabar deposits other metallic minerals are inconspicuous. The iron sulphides, pyrite and the less common marcasite, are almost nowhere completely absent, and in a few de-

[17] Dreyer, R. M.: The problem of variation in cinnabar coloration. Amer. Miner., 26: 195, 1941.

posits they are abundant. Some of the iron sulphide crystallized before and some after the cinnabar. Stibnite, generally in subordinate quantities, has been reported from several localities. At Sulphur Bank in California, small quantities of stibnite are crystallizing from the water of the present hot springs. Copper stains are visible in the ore of several localities, but the copper sulphides are rarely abundant enough to be recognized on inspection. Traces of gold and silver are revealed by assay in many quicksilver deposits but very few contain enough of these metals to be of commercial interest; on the other hand, small quantities of cinnabar and other quicksilver minerals are found in numerous precious metal deposits, both lodes and placers.

Iron oxides are widespread in quicksilver deposits of several different kinds, although commonly in subordinate amount. Finely crystalline red hematite is the most abundant of these and in many places is so intimately mingled with the cinnabar as to indicate that it is of nearly contemporaneous origin. Hydrous iron oxides are also fairly common. In lodes formed close to the surface some of these may have crystallized at about the same time as the hypogene cinnabar. Some iron oxides so closely approximate the appearance of earthy cinnabar as to be difficult to discriminate from that mineral. Millerite, garnierite, and other nickel minerals are widespread in minor quantities in cinnabar lodes associated with altered serpentine. Locally at least they are earlier than cinnabar but lie in veins or coat fractures in the silica-carbonate rock.

Gangue Minerals. The three kinds of introduced gangue minerals most plentiful in cinnabar deposits in this country are the carbonates, different forms of fine-grained silica, and clay minerals. Others are commonly present in small quantities. In parts of most lodes the cinnabar is almost or entirely free from introduced gangue, and in some cinnabar deposits introduced gangue is very sparse throughout. In most places the greater part of the gangue minerals crystallized before most of the cinnabar formed. This generalization, however, is complicated by the fact that in some districts there were recurrent surges of

deposition. In these cinnabar may have been deposited in each of several surges, but mainly during the closing episode of each surge. Commonly most of the cinnabar formed during one of the late stages of the mineralization.

Carbonate minerals are especially plentiful in some of the lodes associated with serpentine or its derivative, silica-carbonate rock, mostly in California; also in those in limestone, best represented by the deposits of the Terlingua region in Texas. The lodes closely associated with serpentine contain dolomite, ankerite, some calcite, and, rarely, siderite, whereas in the Mayacmas district, at least, in those in sandstone of the Franciscan formation the principal carbonate is calcite. In some districts in California north of San Francisco, breccia zones in silica-carbonate rock cemented with coarsely crystalline dolomite are relatively rich in cinnabar. Breccia of this kind appears to be less plentiful in the Californian districts south of San Francisco. In western Oregon early Tertiary rocks have been extensively altered in the vicinity of the quicksilver lodes and the altered rocks contain calcite and siderite. In Texas the dominant gangue mineral in most of the deposits is coarsely crystalline calcite stained yellow with bitumen. The country rock of nearly all the deposits here consists of limestone and calcareous shale and flagstone of Cretaceous age. In southeastern Oregon and other localities aragonite and calcite formed late in the history of some of the opalitic deposits.

Nearly all quicksilver lodes contain introduced silica minerals, mostly fine-grained. In places, such as parts of the Terlingua district, Texas, they are so inconspicuous that it is almost impossible to detect them except through chemical analysis. Chalcedony is probably the most abundant of the silica minerals, but opal and quartz also occur. As quicksilver deposits, like most other metallic lodes, are formed in successive stages, it seems probable that much of the silica deposited at early stages was opal, since dehydrated, recrystallized, and replaced, so that other forms of silica are now more common. Some, possibly a large proportion, of the silica in the present lodes is believed to have come from the neighboring country rocks but to have been so redis-

tributed and so mingled with material from ascending solutions as to obscure its origin. The "twice-silicified reefs" of the Mayacmas district, California,[18] and the silica thought to have been transferred at the Opalite mine, Oregon, from underlying beds now kaolinized,[19] are among the numerous probable examples of this process.

In the Coast Ranges of California the deposits contain halloysite, nontronite, and probably other clay minerals. In most of the Californian mines mapped by the writer clay minerals are rarely abundant but in some, particularly in districts south of San Francisco, they are reported to be plentiful. The "black alta"[20] which forms the hanging wall of many of the ore shoots is a clay gouge impregnated with bitumen. The term "black alta," however, is loosely used and in some places may be applied to any dark, fine-grained material in the hanging wall of a lode. Montmorillonite, halloysite, and kaolinite have been recognized in opalitic and similar deposits and in a few such lodes, particularly near Weiser, Idaho, the better ore consists of opalitic material that was extensively kaolinized before much of the cinnabar was introduced. Dickite is the most distinctive gangue mineral in the Arkansas deposits, although quartz is somewhat more abundant.

Bituminous substances are among the most widespread and diverse of the minor constituents of quicksilver lodes. Many, probably most, of the deposits in the Californian Coast Ranges contain bituminous substances, mostly with paraffin bases. Bitumens are not known in the lodes in sedimentary rocks in southwestern Oregon, but varieties of gilsonite and similar substances are rather plentiful in volcanic rocks in the central part of the state. Asphaltic material is widely distributed and locally abun-

[18] Ross, C. P.: Quicksilver deposits of the Mayacmas and Sulphur Bank districts, California. U. S. Geol. Surv. Bull. 922–L: 333, 1940.

[19] Yates, R. G.: Quicksilver deposits of the Opalite district, Malheur County, Oregon, and Humboldt County, Nevada. U. S. Geol. Surv. Bull. 931, in preparation.

[20] Bradley, W. W.: Quicksilver resources of California. Calif. State Min. Bur. Bull. 78: 155, 1918.

dant in the Texan deposits. Some of it is thickly studded with analcite. Bitumens have not been reported from Arizona nor Arkansas, nor from the typical opalite deposits of Nevada and adjacent states. They are, however, widespread in the lode at Sulphur Bank, California, which has some of the characteristics of opalitic deposits. Methane or some other inflammable substance issues through orifices in the cinnabar-bearing sinter of the Coso district, California. Bitumens are also widespread in foreign deposits of diverse kinds including all of the larger producers,[21] such as those of Idria and Monte Amiata in Italy, Almaden in Spain, some of the lodes in Mexico, Huancavelica in Peru, and others.

The minor constituents of quicksilver deposits in the United States include barite, fluorite, gypsum, sulphur, tourmaline, and zeolites, as well as a number of sulphates and other substances that may or may not be hypogene.

STRUCTURE.

The kinds of structure associated with quicksilver deposits are almost as numerous as the deposits themselves. The one condition that seems common to all is that the structure at the time of mineralization permitted ready passage of solutions that lacked power to make openings for themselves or to move appreciably through the pores of an unfractured rock. Most of the lodes are aggregates of numerous fillings of individually small spaces that were open or under slight pressure at the time of mineralization. In numerous areas the general distribution of the deposits is influenced by major fault systems, but ore shoots appear to have formed more commonly in tension gashes and breccias associated with complementary fractures than in the main fault fissures themselves. Large fault fissures appear generally to have been too tightly compressed or too full of gouge to be suitable sites for cinnabar deposition. It is quite possible, however, that, as some believe, reopening during later adjustments locally has per-

[21] Halse, Edward: Mercury ore. Imp. Inst., Monographs on Min. Res. with special reference to the British Empire, pp. 40–90, 1923.

mitted formation of quicksilver ore shoots along master faults. Where massive rocks are traversed by numerous systematically grouped fractures with negligible displacement, commonly called joints, those so related to regional structure as to have been under tension at the time of mineralization may be lined with cinnabar. Fractures of this and other kinds on anticlines appear to have been favorable loci of deposition, a fact which was long ago recognized by Udden.[22] He regarded these structures as suitable for collecting and impeding ascending fluids and drew an analogy with the accumulation of petroleum in similar structures, although he emphasized that, unlike petroleum, quicksilver deposits are confined to fissures and faulted parts of the anticlines. It seems, however, that the essential condition is that fractures, of whatever origin, are under tension in anticlines. Such fractures tend to gape open when under sufficiently low hydrostatic head.

Any kind of structure so unstable as to be subject to minor adjustments may be suitable as a site for quicksilver deposition because, when such movements take place under low pressures, the resulting fissures tend to remain open. This is one of the reasons so many of the quicksilver deposits in California are close to the borders of serpentine masses in sandstone. These two rocks are so unlike in their reactions to pressure that faults and fracture zones at their borders are common. The serpentine and its derivative, silica-carbonate rock, record repeated hydrothermal alteration accompanied by volume changes (expansion, contraction, or both) that have fissured the enclosing rock and brecciated the silica-carbonate rock. Volume changes have taken place as a result of the successive steps in silicification that characterize the so-called opalite deposits. In these there has clearly been a decrease in volume, and the breccia and fissure zones are within the opalite rather than on its borders.

Many of the fractures that have localized quicksilver deposits are of small displacement. Many have displacements of a few

[22] Udden, J. A.: The anticlinal theory as applied to some quicksilver deposits. Univ. Texas Bull. 1822, 30 pp., 1918.

inches or a few feet. Examples are plentiful along borders of serpentine masses in California, in the limestone of the Terlingua district, Texas, in opalite deposits such as those near Weiser, Idaho, and in other localities.

The major structural features, which largely control the fracture pattern in each area, originated during Mesozoic or earlier time; however, it is believed that the minor fractures (joints) that contain much of the ore were opened or reopened during much more recent adjustments which took place after extensive erosion.

Deposits in which the cinnabar is disseminated through the pores of an unfractured rock are rare. In some lodes that appear to be of this kind the cinnabar lies along tiny fractures, many of them of microscopic dimensions. Some of the ore in apparently massive argillaceous limestone in the Terlingua district and much of that in sandstone of the Franciscan formation at the Oat Hill mine, Napa County, California, is of this character. Disseminated cinnabar is reported from Arkansas, but it "is commonly found only a fraction of an inch from the fractures along which the ore solution probably travelled."[23] Openings, more or less minute, along schistose partings, bedding planes, and similar structures may serve as sites for cinnabar deposition, especially where they have been opened by buckling of the rocks. Such facts support the concept gained from microscopic study of ore from non-disseminated deposits that cinnabar does not extensively replace rock. Most of the cinnabar in the deposits studied fills pre-existing open spaces although the openings have in places been enlarged chemically during cinnabar deposition.

The opalitic deposits are of a special type in which part of the cinnabar is diffused in minute specks through the fine-grained silica that has replaced the original rock. This cinnabar, however, formed in one of the early steps in the process and much of the mineral in material of commercial tenor is in later films along fractures, apparently with only slight enlargement by replacement.

[23] Reed, J. C., and Wells, F. G.: U. S. Geol. Surv. Bull. 886–C, p. 51.

In quicksilver deposits, as in other metallic lodes, any structural condition that helped to check the flow of the mineralizing solutions at one place increased the proportion of material deposited there and, hence, tended to localize ore shoots. This idea has been emphasized by Schuette,[24] who goes so far as to believe that nearly all valuable quicksilver ore shoots are localized under relatively impervious rock that has acted as a dam to the ascending solutions. Such a condition is especially well illustrated by the deposits in Texas where many of the ore shoots are in a zone of a little more than 50 feet in vertical extent close to the contact between the Del Rio clay and the underlying Devils River limestone, in which the size of fracture openings had been greatly increased by ground water solution prior to mineralization. Deposits in California beneath the peculiar bitumen-impregnated gouge locally called "black alta" constitute other good examples. In the New Almaden mine the alta bodies are exceptionally large, and Schuette may well be correct in his concept that these great masses of alta, so favorably located, constitute one of the dominant reasons for the phenomenal orebodies of this great mine. In other mines in the Coast Ranges some of the bodies of black alta thought to have contributed to the formation of ore shoots are small. For example, in the serpentine mass of the Great Western mine, Lake County, some of the sliver-like inclusions of sandstone of the Franciscan formation, now partly or completely altered to black alta, are less than twenty feet in maximum dimension, but Schuette,[25] who was in charge of exploration at the Great Western in 1938, sought these slivers because of the cinnabar that, in places, lay beneath them. A great variety of other forms of structural traps exists. Some consist of minor irregularities in moderately steep conduits, thought to have checked the rising solutions sufficiently to result in concentration of cinnabar beneath them. Many such irregularities are only a few inches or a few feet wide. In most places the cinna-

[24] Schuette, C. N.: Occurrence of quicksilver ore bodies. A. I. M. E. Trans., 1931, page 406, 1931.

[25] Schuette, C. N. Personal communication.

bar is in rock that, for one reason or another, was more permeable than the adjoining rock at the time of mineralization. In many such situations, especially where contacts are steep, it is a question whether the permeability of the solution channel or the trapping effect of the tighter material nearby was the dominant influence in localizing the ore.

In many places the permeability of the ore channels results from the presence of numerous closely spaced fractures in a rock that is fine-grained and tightly cemented but distinctly more brittle than neighboring rocks. The silica-carbonate rock derived from serpentine, and the chert beds in the Franciscan formation, both in California, the diabase dikes of the Bottle Creek district, Nevada,[26] and the silicified rock of opalite deposits in numerous districts are examples of such rocks. Commonly the fractures, and consequently the cinnabar films in the brittle rocks, are irregular and rather unsystematic, both in size and distribution.

The permeability of the passages followed by solutions influences deposition in several different ways. The ascending solutions may be confined, almost completely, to relatively permeable zones, so that the wall rocks are scarcely affected by the mineralization. Where a certain quantity of solution encounters a zone in which the openings available for passage are increased, velocity is decreased and opportunity for deposition of the material in solution is improved, in accord with Thiem's formula [27] which states that the quantity of water equals the porosity of the water-bearing material, times the cross sectional area of the conduit, times the velocity. Some of the structural features that have been regarded as traps tend to add to the openings available either by increase in shattering of the rock or of the cross-sectional area of the channels by bulging of the walls. Relatively flat places along a fault and some of the irregularities in the

[26] Roberts, R. J.: Quicksilver deposits of the Bottle Creek district, Humboldt County, Nevada. U. S. Geol. Surv. Bull. 922–A: 13–14, 1940.

[27] Wenzel, L. K.: The Thiem method for determining permeability of water-bearing materials. U. S. Geol. Surv. W. S. P. 679–A: 6, 1926.

hanging walls of channels of different kinds may be in this category. A barrier across a solution channel would decrease the velocity of the solution, but if sufficiently effective to produce stagnation, would tend to prevent formation of an ore shoot because fresh supplies of solution could not reach the area beneath such a trap. Thiem's formula and most of the statements made in the present paragraph apply strictly only if the velocity of the solution is low. Under some conditions, particularly if open spaces in the channels are exceptionally large, velocities are high enough to produce turbulent flow similar to that which occurs in ground water circulating through large openings.[28] This would result in more irregular distribution of the resulting cinnabar.

THE MINERALIZING SOLUTIONS.

Little need be added at present to discussions already in the literature in respect to the chemical character of the solutions of magmatic origin in which the quicksilver is brought up. Dreyer [29] has summarized available information, with citations to the principal papers on the subject, together with observations of his own. His paper is a valuable compilation, but, as already noted,[30] contains some generalizations that should be read with caution, even though his views have been amplified in a later paper.[31]

In brief, the present view is that the hydrothermal solutions that transport mercuric sulphide consist essentially of alkaline sulphides. Precipitation is thought to take place at temperatures below 150° C. and at pressures close to atmospheric as a result of such factors as cooling, and acidification. Dreyer believes that "increased acidity is not generally a common cause of the precipitation of mercuric sulphide," but, as brought out in the

[28] Tolman, C. F.: Ground Water. McGraw-Hill Book Co., New York, 1937, pp. 191–200.

[29] Dreyer, R. M.: The geochemistry of quicksilver mineralization. ECON. GEOL., 35: 17–48, 140–157, 1940.

[30] Fahey, J. J., Fleischer, Michael, and Ross, C. P.: The geochemistry of quicksilver mineralization. Disc., ECON. GEOL., 35: 465–470, 1940.

[31] Dreyer, R. M.: The geochemistry of quicksilver mineralization. Disc., ECON. GEOL., 35: 905–909, 1940.

discussion of his paper, this conclusion is questionable. The common belief that in hydrothermal processes metals are transported in dilute alkaline solutions has recently been challenged by Garrels,[32] because he thinks that, theoretically, "metal sulphides reach their minimum solubility in such an environment." He notes, however, that quicksilver is one of the few metals truly soluble in alkaline solutions, so that his objections would not apply to this metal. Direct evidence that solutions of the alkalis have traversed the rocks is absent in many quicksilver mining districts, perhaps because so many of the salts of the alkalis are highly soluble. Possibly the Terlingua district, Texas,[33] constitutes an example of retention in the rocks of the alkali in the mineralizing solutions. Here soda-rich zeolites are abundant in the igneous rocks and a few of the lodes. Also the feldspars of the igneous rocks have been extensively albitized late in their history.[34]

The mineral content of most cinnabar deposits suggests that the solutions that brought the quicksilver were of simple composition. Perhaps the principal bases were sodium and quicksilver and the acid radicles were mainly sulphide and silica, with minor amounts of other constituents. Locally, at least, as suggested below, sufficient other material was added from ground water prior to mineralization to modify markedly the resulting assemblage of gangue minerals.

RELATIONS TO WATER OF METEORIC ORIGIN.

Most cinnabar deposits appear to have been deposited within depths ranging perhaps to 3,000 or 4,000 feet below the then-existing surface, and some formed at or immediately beneath the surface. Opalitic and related lodes have so many of the features characteristic of hot-spring deposits as to indicate that they came from water essentially like that of hot springs. Evidence is ac-

[32] Garrels, Robert: Vein-forming solutions. ECON. GEOL., 36: 663–665, 1941.

[33] Ross, C. P.: The quicksilver deposits of the Terlingua region, Texas. ECON. GEOL., 36: 39, 1941.

[34] Stocking, H. E.: The process of analcitization in the Terlingua, Texas, region. Dissertation, Johns Hopkins Univ., 1936.

cumulating that most of these deposits, contrary to beliefs formerly held, formed beneath the surface of the ground. At some hot springs still issue, but proof is lacking that any of the present hot springs are depositing significant quantities of quicksilver minerals.

Tests of water from Sulphur Bank, California, recently made in the laboratory of the Geological Survey fail to show quicksilver, and it seems probable that part or all of that reported in previous analyses was in the form of suspended cinnabar particles in the water. At Steamboat Springs, Nevada, the other locality most commonly cited in this connection, the present cinnabar deposits were all formed in highly altered bedrock rather than in sinter deposited at the surface, as is evident from unpublished studies by Ralph Roberts and the writer. Becker [35] reports traces of mercuric sulphide in spring water and spring deposits in this area, but further studies are required to show that this is brought up in solution rather than in suspension. It is clear, however, that many of the opalitic deposits formed within a few score or at most a few hundred feet of the surface and in a few, such as those of the Coso district, Inyo County, California, the cinnabar crystallized in hot spring deposits at the surface. Studies of this deposit by R. G. Yates and the writer [36] show that minute quantities of quicksilver are brought up in some of the present emanations, but in an undetermined form. The relation between the surface and other kinds of cinnabar deposits is less close. Some deposits, however, such as those in limestone caverns in Texas, were clearly controlled in position by their relations to the then-existing surface, and the textural features of many show that they crystallize at such shallow depths that pressures were not high.

The features outlined above suggest that in many districts the quicksilver solutions penetrated into and even passed upward

[35] Becker, G. F.: Geology of the quicksilver deposits of the Pacific Slope. U. S. Geol. Surv., Mon. 13: 342–353, 1858.

[36] Ross, C. P., and Yates, R. S.: The Coso quicksilver district, Inyo County, California. U. S. Geol. Surv. Bull. 931, in preparation.

through the zone of ground water saturation. Graton[37] is of the opinion that where ascending hydrothermal solutions pass upward into the region of free subsurface water they have sufficient volume and pressure to flush out and displace this water from the channels utilized by the solutions, holding it at a distance until hypogene mineralization is virtually complete. This concept seems the most probable one with respect to the genesis of most metallic lodes but is apparently not applicable to some at least of the cinnabar deposits. These may have formed under insufficient pressure to permit the flushing out of ground water from their conduits. It is now generally accepted that much of the volume of the water of hot springs is of meteoric rather than magmatic origin. The solutions that deposited cinnabar at and near the surface seem clearly to have resembled hot-spring waters in this as in other respects. There may have been all gradations between quicksilver-bearing waters of this character and those that, as Graton visualizes, may have penetrated the zone of saturation but did not mingle with ground water. The cinnabar lodes in the Coast Ranges in California are thought to be among those that show the influence of mingling of ascending solutions with ground water. There is little evidence in most of these deposits that the solutions attacked the wall rocks extensively. Nevertheless the gangue minerals of the cinnabar lodes vary within short distances in a manner that seems best explained by mingling of the ascending solutions with ground water that varied in composition in accord with the character of the nearby rocks. Where magnesian minerals are plentiful in the country rocks, ankerite and even magnesite appear in the lodes. Where the original serpentine is largely silicified, dolomite is prevalent, and in sandstones remote from serpentine calcite is present. Most of the cinnabar was formed after the carbonates had crystallized. Another example is afforded by opalitic deposits in Nevada and Oregon which are thought to have derived their silica from leaching of the rocks below.

[37] Graton, L. C.: The depth-zones in ore deposition. ECON. GEOL., 28: 520, 1933.

In several districts it is evident that mingling of rising solutions, themselves probably composed in part of ground water, with meteoric water from relatively close to the surface has aided in precipitation of cinnabar and associated minerals. This matter is discussed in a brief summary of studies made in the Terlingua region, Texas, in 1934.[38] At Sulphur Bank, California, most of the production has come from the lower part of a Recent basalt flow. The upper part of this flow has been thoroughly opalized by water that contained sulphuric acid, produced by oxidation of ascending hydrogen sulphide by air brought down in rain water percolating from the surface of the low mound formed by the lava flow. Much of the cinnabar is localized in a zone of narrow vertical range in which the rising solutions met the acid water from the upper part of the flow. This descending water was still well above the water table and would have been more acid than the ground water even if the sulphuric acid mentioned above had not been added to it. Similarly Yates [39] believes that ore shoots at the Opalite mine were localized at the horizon where ascending solutions encountered a sheet of vadose water that was moving down the dip of slightly inclined lake beds toward the zone of saturation. Water from the immediate vicinity of the surface would be, presumably, more acid and during most of the year cooler than the ground water. Both coolness and acidity would aid precipitation.

Obviously lodes formed as indicated above, in part under the influence of oxygenated water from the uppermost zones of circulation of water under ground, will have characteristic mineralogic features that ally them with parts of lodes of other kinds that have been altered after erosion has brought them into these upper zones. Under such conditions the conventional differences between supergene and hypogene processes and minerals become difficult to apply. The presence of oxides, even hydrous oxides,

[38] Ross, C. P.: The quicksilver deposits of the Terlingua region, Texas. ECON. GEOL., 36: 137–141, 1941.

[39] Yates, R. G.: Quicksilver deposits of the Opalite district, Malheur County, Oregon, and Humboldt County, Nevada. U. S. Geol. Surv. Bull. 931–N, in press.

in a quicksilver deposit is not necessarily proof that supergene processes, in the ordinary sense, have been operative. Cinnabar itself is a resistant mineral and there is reason to believe that other quicksilver minerals, such as metacinnabar, the native metal, and the oxychlorides, regarded as supergene by many observers, may be mainly products of different stages in the original mineralization. Some of the reasons for this belief have been outlined in previous papers.[40] Final conclusions on this matter await more complete information, but the belief is growing that redistribution of quicksilver as a result of the action of descending meteoric water distinct from that relative to original mineralization has taken place in numerous localities but generally on a scale too small to be of economic significance. Some small but workable deposits that seem clearly to be of supergene origin are known.

Cinnabar is so friable that specks of it small enough to be easily carried in water or vapor are easily formed. Conspicuous but economically trivial films of cinnabar on mine walls appear to have been produced in this way. Similarly mechanical transportation of tiny, solid particles may furnish the basis for some, at least, of the reports of quicksilver carried in modern hot spring emanations.

CONCLUSION.

The facts and inferences outlined above indicate that quicksilver deposits of the types mined in the United States are formed at geologically shallow depths and at relatively low temperatures and pressures. Dilute hydrothermal solutions of relatively simple composition commonly are modified by mingling with ground water before precipitation occurs. As a result the gangue of the deposits varies in composition in accord with the local environment. The fact that few other metals are associated with cinnabar deposits and that such deposits tend to be remote from those

[40] Ross, C. P.: Quicksilver deposits of the Mount Diablo district, Contra Costa County, California. U. S. Geol. Surv. Bull. 922–B: 40–44, 51–52, 1940. The quicksilver deposits of the Terlingua region, Texas. ECON. GEOL., 36: 132–133, 139–142, 1941.

of other metals suggests that cinnabar lodes may be end-products formed after any other constituents that may have been in the hydrothermal solutions when they left their magmatic source had been precipitated in the course of earlier reactions. Some of the deposits that contain quicksilver minerals other than cinnabar, particularly those with quicksilver-bearing tetrahedrite, may have been formed at significantly greater depths than the typical cinnabar lodes but too little is known about these for decision on such matters as yet.

Cinnabar ore is confined to parts of the rock where the solutions found easy access and tends to be concentrated where suitable structural traps existed. Precipitation of cinnabar is caused by any changes that decrease the temperature and alkalinity. The factors dominant in precipitation differ with the local conditions, but decrease in alkalinity may be one of the principal factors in most deposits. In some places, at least, precipitation is aided by encounter with cool water descending from above the water table. Such water even where not actually acid, is so low in alkalinity that dilution by it promotes precipitation. Supergene changes have taken place in some cinnabar lodes, but so far evidence is lacking to prove that in most deposits such changes, distinct from late steps in original deposition, have been on a sufficiently large scale to be of economic significance.

WASHINGTON, D. C.,
July 6, 1942.

A STUDY OF THE GEOLOGY AND ORE DEPOSITS OF THE ASHBROOK SILVER MINING DISTRICT, UTAH.[1]

VICTOR E. PETERSON.

ABSTRACT.

The ore of the Ashbrook district occurs as irregular patchy replacement deposits in fractured and folded limestone, whose structure has strongly influenced distribution of the orebodies. The distribution is controlled by small crenulations within the limestone, the apical portions of which are usually marked by a strong development of secondary calcite, which is generally more susceptible to replacement by mineralizing solutions. Due to two episodes of minor fracturing, the sequence of primary mineralization can be devided into three periods of deposition: (1) rhodochrosite to the exclusion of other minerals, (2) quartz, pyrite, arsenopyrite, sphalerite, and minor amounts of chalcopyrite, and (3) galena, argyrodite, pyrargyrite, pearceite, tennantite, and an unidentified mineral. The second phase seems to be responsible for the minor gold values in the ores. Processes of enrichment and oxidation have produced a large group of secondary minerals. Alteration of the primary silver minerals has been mainly to argentite and native silver which together account for the main values found in the ores.

THE Ashbrook district, one of the oldest mining districts in Utah, has at times attained a very high-ranking position in the State's total silver production. Information regarding the nature of the geology and ore deposits is very limited in the literature and it is hoped that the present paper will add to the knowledge of this area.

The district is situated in the extreme northwestern corner of Boxelder County, approximately 10 miles east of the Nevada State line and 1 mile south of the Idaho State line, and comprises 71 claims and at least 7 mines which have at one time or another been of considerable local interest. In all, there have been ap-

[1] The material presented here is taken from the writer's doctor's dissertation on file at the University of Chicago.

proximately 5 miles of development work completed, of which more than 90 per cent is accounted for in one mine.

General Geology.

Physiography. The district is situated high on the west side of the Goose Creek Range, approximately within the area where this range merges with a northwestward extension of the Grouse Creek Range on the east. The topography of the district, similar to that of the higher reaches of many ranges in this part of the Great Basin, consists of high, smooth, east-west transverse ridges of nearly equal elevation, separated by broad, roughly V-shaped valleys, which descend gently from the high, northward-trending crest of the range. The total relief in the district is nearly 2,000 feet, and the higher points attain elevations of about 8,000 feet. Probably the most outstanding features of the topography in the area are the marked accordance in elevation and slope of the principal east-west ridges, and their accordance in elevation with the many local flattened areas. This condition prevails over a large part of the western slopes of the Goose Creek Range in Utah. Conclusive evidence indicates that this striking accordances is an expression of a once continuous erosion surface which correlates closely in time with the Snowdrift peneplain defined by Mansfield [2] in southeastern Idaho. A similar erosion surface has been recognized by Anderson [3] in eastern Cassia County, Idaho, approximately 20 miles northeast of this district. The east-west valleys cutting this old erosion surface appear to owe their origin primarily to a period of erosion closely corresponding to that responsible for the Tygee erosion surface in southeastern Idaho, also defined by Mansfield.[4] The chief evidence for these conclusions is the relationship which exists between the local topography and the Salt Lake formation, which is known to be deposited on the Tygee erosion surface in southeastern Idaho.

[2] Mansfield, G. R.: Geology and mineral resources of part of southeastern Idaho. U. S. Geol. Surv., Prof. Pap. 152: 8, 11, 1927.

[3] Anderson, A. L.: Geology and mineral resources of eastern Cassia county, Idaho. Idaho Bur. Mines and Geol. Bull. 14: 9, 1931.

[4] Mansfield, G. R.: *op. cit.*, p. 15.

Structural Geology. The most prominent structure within the district is a broad, spoon-like syncline which plunges at a low angle to the south. Except for the complicated structures within the pre-Cambrian rocks, all other major structural features in the district are superimposed on this structure. The district is

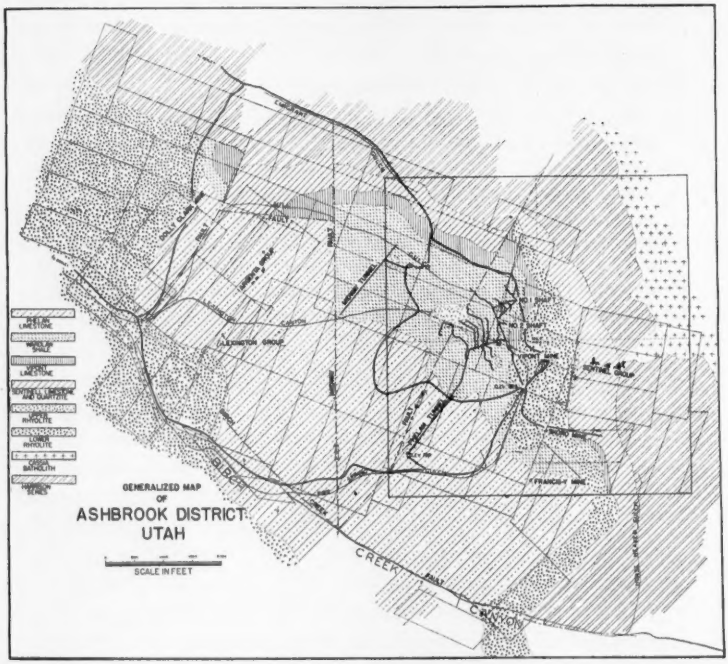


FIG. 1. The Ashbrook District, Utah.

situated in the north end of this fold (Fig. 1). The length of the fold is not known, though it is apparent that it extends several miles beyond the southern limits of the district. Prior to extensive displacement by north-south and east-west normal faults, the fold appears to have been approximately symmetrical with the steepest dips on the flanks about 20 degrees. The fold is not

simple, but is complicated by the presence of numerous major and minor crenulations, the largest of which have a roughly radial relation with respect to the major fold. These crenulations have played a very important part in the localization and distribution of the ore deposits of the district. Further complications in the general outline of the fold have been introduced by the warping associated with two periods of magmatic intrusion.

Though on a large scale, faulting within the district is not complicated and is confined mainly to several large normal faults. These faults may have played an important part in the distribution of the intrusives associated with the ore deposits, but no evidence was observed that would indicate that they directly influenced the distribution of the deposits themselves. The faults all appear to belong to approximately the same period of deformation and antedate both the late intrusives and the ore deposits. It is assumed from physiographic evidence, from which the age of the late intrusives can be roughly estimated, that these faults belong to a period of deformation closely following the Laramide revolution, when compressive forces had ceased, and tensional forces prevailed. It is clear that these faults are much older than the late "Basin-Range" faulting. Faults of similar age have been described by Anderson [5] in eastern Cassia County, Idaho, and are attributed to deformation closely following the Laramide revolution. The formation of the major fold of the district appears to have been concurrent with Laramide disturbances.

Stratigraphy. Within or in the immediate vicinity of the district there are rocks ranging from pre-Cambrian to late Tertiary. They include a series of metamorphosed pre-Cambrian schists, slates, and gneisses with beds of marble and quartzite, overlain by Paleozoic limestones and shale. The lower part of the series, including the Paleozoic rocks, has been intruded by a large granite batholith and its associated pegmatites, which Anderson [6] considers to be of late Cretaceous or early Eocene age. Rhyolite porphyries intrude the upper part of the series in a sill-

[5] Anderson, A. L.: *op. cit*, pp. 78 and 87.
[6] *Ibid.*, p. 55.

like fashion. These later intrusives appear to belong to an earlier period of volcanic activity than the quartz latites of eastern Cassia County, Idaho, possibly the late Eocene or early Oligocene. The Salt Lake formation of late Tertiary age (Pliocene?) crops out just off the western edge of the district.

Definition of the recognized local stratigraphic units is difficult, due to their remoteness from described stratigraphic sections and the scarcity of fossils in the Paleozoic rocks. Correlation was made here entirely on lithologic similarities between the local series and defined stratigraphic units within the surrounding region.

Pre-Cambrian. The pre-Cambrian rocks of the district are equivalent to part of the great thickness of contorted and metamorphosed rocks which Anderson [7] has described in eastern Cassia County, Idaho, and named the Harrison series from their occurrence at Harrison Peak in the Albion Range. It is not known how much nor which member or members of this three-fold series, are represented in the Ashbrook district. The lithology of the local series most closely resembles parts of the middle and upper divisions.

Paleozoic. The Paleozoic rocks of the district are divided into five separate lithologic units, the general features of which are outlined in Table 1. The lithology of the two lowest members of this series closely resembles rocks of Cambrian age in the Raft River Range, approximately 30 miles east of the district. The lowest quartzite member closely resembles parts of the Brigham quartzite in eastern Boxelder County, and they are tentatively correlated. The Sentinell limestone is the host rock for the deposits of the Skoro and Sentinell mines.

In eastern Cassia County, Idaho, Anderson [8] has described a series of rocks of Carboniferous age which are very similar to the three highest members of the local Paleozoic series, and it is tentatively considered that the two series are correlatives. The two highest limestone members of the local Paleozoic series make

[7] *Ibid.*, pp. 24–27.
[8] *Ibid.*, p. 31.

up the host rocks for the ore deposits of the Vipont and old Lexington-Argenta mines.

Igneous Rocks. The igneous rocks of the district can be divided into two groups, the granites and pegmatites of the Cassia batholith, and the later rhyolite porphyries. Since it was apparent during the investigation that the first of these groups has

TABLE 1.

GENERALIZED STRATIGRAPHIC SECTION OF ASHBROOK DISTRICT.

Age.		Formation.	Thickness.	General Lithologic Character.
Paleozoics	Carboniferous (?)	Phelan limestone	1,000–1,500	Coarse to finely crystalline, light blue-gray, massive bedded limestone, locally cherty.
		Wardlaw shale	400	Black to gray, thin to massive bedded, pyritic shale, locally containing lenses of gray dolomitic limestone.
		Vipont limestone	50–100	Fine-grained, crystalline, blue-gray limestone, locally containing numerous small calcite stringers.
	Cambrian (?)	Sentinell limestone	50–100	Light gray, medium grained, crystalline, thin-bedded limestone, usually broken and containing numerous calcite stringers.
		Sentinell quartzite	100–100	White to tan, sucrose quartzite.
Pre-Cambrian	Harrison Series	Upper?	300–400 500–1,000	Usually light gray, contorted and fractured limestone and marble. Massive, white to pinkish, somewhat schistose, vitreous quartzite.
		Middle?	400–600	Fractured and contorted limestone (with mica and marble) interbedded with slates and quartzite.

had no direct relation to the ore deposits, time was not given to a detailed study of those rocks.

The intrusive rocks later than the Cassia batholith are locally referred to two groups, those belonging to the lower or coarse grained porphyry, and those belonging to the upper or fine grained porphyry. Of these, the lower porphyry sill is the most extensive, and crops out in a wide belt almost completely encircling the district. The upper porphyry occurs mainly as small pipes or dikes cutting across the bedding planes of the Wardlaw shale and Vipont

limestone, though locally it assumes a sill-like form. Lithologically, the lower porphyry is a fine grained, even textured rock, containing phenocrysts of quartz and feldspars from 5 mm. to 15 mm. in diameter. It is usually light gray, but often possesses a distinct greenish hue. The upper porphyry is a light gray to grayish tan, fine grained rock, containing phenocrysts of quartz and feldspars from 2 mm. to 5 mm. in diameter. The phenocrysts of quartz are predominantly larger than those of the feldspars. Except for textural differences the two rocks are similar, the primary minerals of both including andesine, quartz, and sanidine, with minor amounts of magnetite and apatite. Secondary minerals are chlorite, calcite, white mica, and leucoxene. Quartz, sanidine, and andesine make up the phenocrysts in both rocks. The fine grained matrix is composed mainly of these same minerals in about equivalent proportions. These rocks are quartz-sanidine rhyolite porphyries. They represent different phases of the same intrusion, and though conveniently referred to two groups, are strictly speaking of one rock type, differing essentially only as to texture and mode of occurrence.

ORE DEPOSITS.

General. All the ore deposits thus far found in the district are replacements in limestone and all have an irregular, patchy distribution, whether oxides or sulphides. Deposits are similar in shape and form in each of the three limestone formations recognized in the district, but deposits in the Vipont limestone are much more abundant and widespread. The irregularity and bunchy distribution of the ores has in part accounted for the delayed and sporadic development of the district. It has never been possible to block out ore with any success in any of the mines, nor to predict the character of ore or size of a single, small orebody.

The distribution of the mines and prospects within the district is clearly related to the distribution of the late intrusives (Fig. 1). Extensive prospecting has revealed that ore values are almost invariably associated with limestone in close proximity

to the porphyries and that the values in general diminish with distance from these rocks. Replacements have not been general, but in all cases have been localized by structural features within the limestones. Most notable of these features are the small crenulations in the limestone, along the bedding planes of which abundant secondary calcite has developed.

Accurate information as to the production of each of the mines in the district is not available; statistical reports give only the district production with no apportionment to the various mines. From available information, however, it is evident that the Vipont mine alone has yielded more than 95 per cent of the district's total production.

Production from 1873, when the first claim was staked, to 1899 aggregated approximately seven hundred tons with a value of twenty thousand dollars. Production from 1899 to 1917 has been summarized in the following table by Heikes:[9]

PRODUCTION FROM ASHBROOK DISTRICT 1899–1917.

Year.	Ore Mined.	Gold.		Silver.		Total Value.
	Short Tons.	Ounces.	Value.	Ounces.	Value.	
1899	225	37.17	$ 768	18,245	$10,947	$11,715
1900	22	9.04	187	3,374	2,092	2,279
1901	200	3.30	68	1,639	983	1,051
1902	28	5.58	115	1,640	869	984
1903	91	7.75	160	3,189	1,722	1,882
1904	119	4.30	89	1,612	923	1,012
1917	27	3.74	77	1,400	1,153	1,330
	*712	70.88	$1,464	31,099	$18,869	$20,153

* Also yielded 125 tons of concentrate.

During the period 1905 to 1917, work done in the district was principally for the purpose of holding claims, and the Midway tunnel and A-tunnel of the Vipont mine were completed. Active large scale development work in the district did not start until 1918, when the Vipont Silver Mining Company was formed by

[9] Heikes, V. C.: Ore deposits of Utah. U. S. Geol. Surv. Prof. Pap. 111: 495, 1920.

a New York exploration company. From 1918 until 1923, when this company ceased exploration work, the district attained a high ranking position in the State's total production of silver. Production during these years is summarized below.

PRODUCTION FROM ASHBROOK DISTRICT 1918–1923.

Year.	Gold.		Silver.	
	Ounces.	Value.	Ounces.	Value.
1918	116.10	$ 2,400	21,009	21,009
1919	———	———	———	———
1920	1,024.27	21,188	525,853	573,180
1921	1,331.00	27,527	716,808	716,809
1922	2,976.03	61,520	1,043,671	1,043,761
1923	435.38		217,534	

The Vipont Silver Mining Company discontinued active operations on August 1, 1923, when buying under the Pittman Act ceased and the price of silver dropped from $1.00 to $0.64 an ounce. Production from this date until 1929, when the mining and milling machinery was sold and moved away from the mine, was small.

The mines of the district remained idle until 1934, when operations at the Vipont mine were renewed on a small scale. These operations have consisted mainly of mining small bodies of high grade ore and have proceeded sporadically to the present date. The total production from 1934 to the present has been about 300,000 ounces of silver and several hundred ounces of gold.

The Vipont Mine.

The orebodies of the Vipont mine occur in the Vipont limestone, between the overlying Wardlaw shale and underlying lower porphyry. They are not confined to a particular horizon within this formation, but are distributed from the porphyry footwall to the shale hanging wall, and usually conform to bedding planes (Fig. 2). Structural features within the limestone, though not clearly discernible in all cases, appear to have been the most im-

portant factors in localizing the ore deposits. The workings of the mine are along the northeastern flank of the synclinal fold forming the major structure of the district. Reflection of this fold is seen in the outline of the main levels of the mine; these main levels correspond in general to the contact between the Vi-

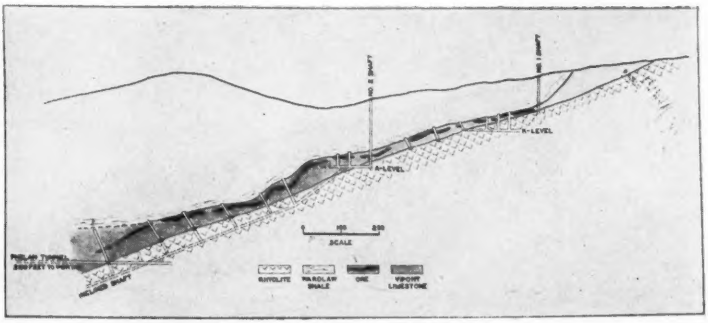


FIG. 2. Section *A–A'* through the Vipont mine showing general distribution of orebodies through the Vipont limestone.

pont limestone and the lower porphyry (Fig. 3). As shown above, this major fold is not simple in form. Superimposed upon it are numerous small crenulations, the largest of which have a roughly radial distribution with respect to the major fold. Close examination of the stope map of the mine reveals that the main stopes also have a roughly radial distribution with respect to the major fold. The main "stope runs" correspond to broad, complicated, anticlinal crenulations on the major fold. The intermediate areas between the main "stope runs" appear to represent the synclinal crenulations. Minor crenulations, together with small irregular slips, complicate these general structural features. Associated with the formation of these crenulations there has locally been development of much secondary calcite in stringers usually paralleling the bedding planes. Development of this calcite has been most extensive in the crests of the crenulations and ordinarily pinches toward the flanks. It should be under-

stood that these crenulations are not neceessarily symmetrical, but show great variation in size, shape, and continuity.

In conjunction with these crenulations, the structure is further complicated by minor slips and localized zones of fracture.

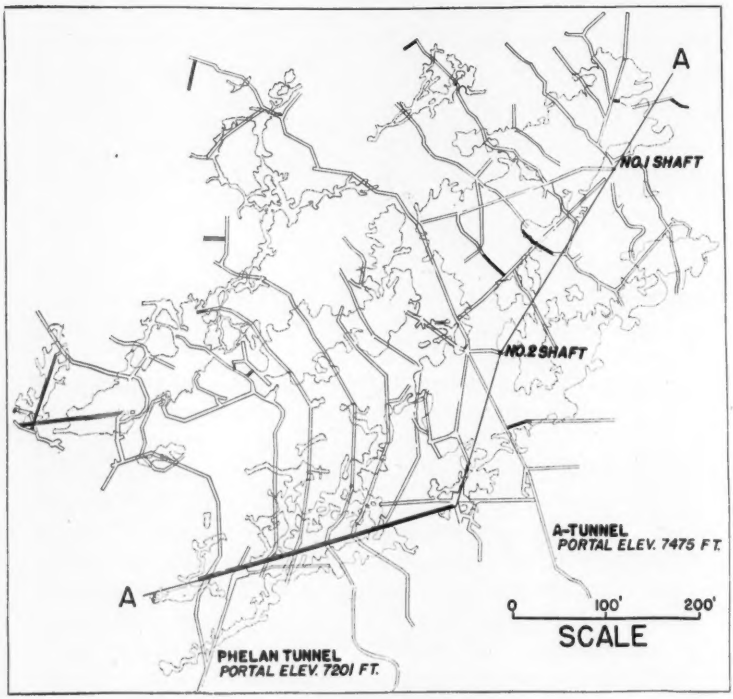


FIG. 3. Generalized stope map of Vipont mine.

These zones of fracture, like the apical portions of the crenulations, in which there has been extensive development of secondary calcite, have formed the most vulnerable areas of replacement by mineralizing solutions. Their intricate form and complex relation to each other account to a large extent for the great variations in the size and continuity of the orebodies.

However, replacement by ore-bearing solutions is not limited to these fractured areas. At present, it is not possible to predict the extension of particular fractured zones beyond known orebodies, or to determine in advance their relation to each other, though it seems possible that through detailed study this might be accomplished.

Size and Form of the Ore Deposits. The ore occurs generally in small patches and pockets of irregular form, connected by small stringers. Commonly they are elongated parallel to the bedding planes in the limestone, and have a roughly lenticular form across their shorter dimensions. The size of the orebodies

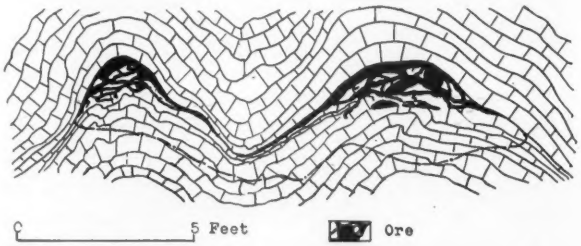


FIG. 4. Idealized sketch showing the relations which existed in two adjacent stopes in the "trespass" above *A*-level.

is extremely variable. An orebody several feet in diameter may in the course of a few feet connect with a scarcely visible stringer; this stringer may again expand into a sizable orebody. For the description of this relation of one orebody to another, the local expression " rolls " is used. These " rolls " correspond to minor crenulations in the limestone, and pinching of the ore is closely associated with the pinching of calcite stringers on the flanks of these crenulations. This relationship is clearly revealed by recent developments in the "trespass" above A-level (Fig. 4), where a small stringer of ore was followed down the flank of an elongate crenulation through almost barren rock into another orebody essentially paralleling the first.

The orebodies are most commonly parallel to the bedding planes of the limestone, but locally, as in Upper K–46 raise, the ore cuts across the bedding planes in a vertical pipe (Fig. 5). At the top of this pipe, the ore spreads out like an umbrella, the upper surface of which conforms to undulations in the limestone. In this case it appears that the pipe acted as a feeder to the upper

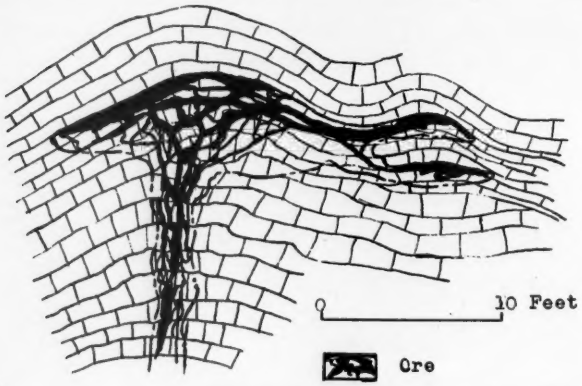


FIG. 5. Idealized sketch showing the relation of ore to the limestone in upper K–46 raise.

part of the orebody and that the resistant bed forming the hanging wall acted as a dam to the rising mineralizing solutions.

The above description has been primarily concerned with deposits of sulphide ore. Though the deposits of oxide ore are not as clearly defined, they appear to have corresponding relationships with the enclosing limestone.

Mineralogy of the Vipont Ores.

Hypogene Minerals. The minerals that are considered primary or hypogene in origin are those formed prior to or overlapping with the deposition of galena, which is apparently the latest of the primary minerals. The primary minerals include: rhodochrosite, sericite, quartz, pyrite, arsenopyrite, gold, spha-

lerite, chalcopyrite, argyrodite, tennantite, pearceite, pyrargyrite, an unidentified mineral, and galena.

Argentite was considered of primary origin in these ores by Vhay [10] and by Foss,[11] who states: "Argentite is probably the essential primary silver mineral in these ores. Its most common occurrence is between quartz grains in the highly silicified limestone, but also filling minute fractures in the same grains." This association of argentite with quartz is not necessarily indicative of primary origin. The writer's observations have revealed no relations between argentite and known primary minerals which are certainly indicative of hypogene origin. Contrary to this, the preponderance of evidence is strongly in favor of a supergene origin for this mineral.

Rhodochrosite is a minor component of the Vipont ores, though locally, it is abundant. Specimens examined came from A–33 raise, where rhodochrosite comprised one of the principal introduced gangue minerals. It occurs in fractured limestone which is shot through by numerous stringers of calcite. Rhodochrosite has replaced both the massive limestone and the later calcite, but is much more extensively developed in the latter. In color it is usually the characteristic pink, though some specimens are markedly bleached. Much of the rhodochrosite is minutely vuggy, the vugs ordinarily being lined with well-formed, clear quartz crystals. In thin section or polished section, quartz is also seen to traverse the rhodochrosite in small irregular veinlets. It is notable that the quartz that forms the smallest veinlets is clear like that which lines the vugs, and that the quartz that constitutes the principal veins is dark in color. This color appears to be due to inclusion of a material resembling manganese dioxide. Here, as in most places, sulphides are more commonly associated with the dark quartz. The change in color from one quartz to the other is transitional and does not represent two separate

[10] Vhay, J. S.: Unpublished professional report on the Vipont mine, March 23, 1937.

[11] Foss, A. L.: The geology of the Vipont mine. Unpublished undergraduate thesis, Univ. Minn., Dept. Geol., 1923, p. 18.

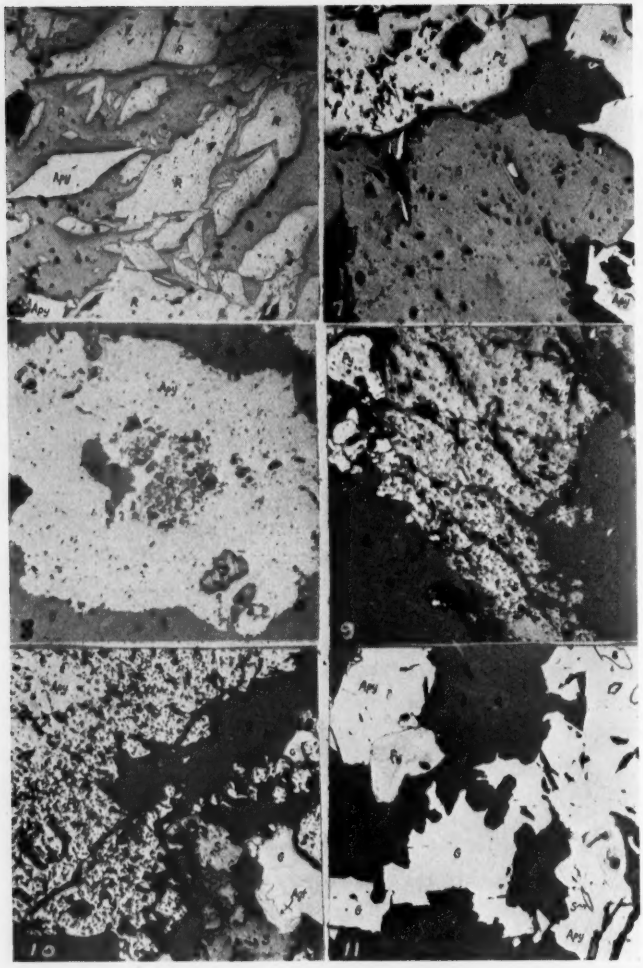


FIG. 6. Quartz (*Q*) filling fractures in rhodochrosite (*R*) and containing automorphic inclusions of arsenopyrite (*Apy*) and rhodochrosite. ×45.

periods of deposition. Pyrite and arsenopyrite have not been observed in contact with rhodochrosite to the exclusion of quartz. These minerals are commonly found in the dark quartz occurring either at some distance from the main masses of rhodochrosite or occupying the darker central portions of quartz veins traversing the rhodochrosite. Pyrite and arsenopyrite are considered to be of a later phase in the sequence of mineralization than rhodochrosite and probably formed subsequent to the development of the first clear quartz.

The relationship between quartz and rhodochrosite is of special interest (Fig. 6). It is noted that quartz traverses the rhodochrosite in irregular veinlets, indicating that quartz is younger than the main masses of rhodochrosite. The edges of these veins show abundant automorphic outlines of rhodochrosite against quartz, and automorphic crystals of rhodochrosite are common inclusions within the quartz veinlets. Automorphic outlines of quartz crystals against rhodochrosite have not been observed. Larger masses of rhodochrosite characterictically show a marginal development of younger rhodochrosite in crystallographic continuity. Small veinlets of later rhodochrosite fill fractures, with slight replacement, in early rhodochrosite, but have never been observed to traverse quartz. From these relationships it is seen that in a strict sense, the quartz veinlets, though definitely later than the rhodochrosite, do not represent

FIG. 7. Pyrite (*Py*), arsenopyrite (*Apy*), and sphalerite (*S*) intergrown with quartz (*Q*). Note automorphic outline of quartz against pyrite and sphalerite at (*X*). ×45.

FIG. 8. Inclusions of quartz (*Q*) and sphalerite (*S*) in arsenopyrite (*Apy*). ×80.

FIG. 9. Quartz (*Q*) filling fractures in sphalerite (*S*). ×90.

FIG. 10. Quartz veinlet (*Q*) traversing pyrite (*Py*) which is intergrown with arsenopyrite (*Apy*). Galena (*G*) interstitial in quartz and sphalerite (*S*) being replaced by argentite (*Agt*). ×80.

FIG. 11. Galena (*G*) interstitial in quartz (*Q*) which conforms to the automorphic outlines of pyrite (*Py*) and arsenopyrite (*Apy*). Arsenopyrite containing inclusions of pyrite and with fractures filled by quartz and sphalerite (*S*). ×80.

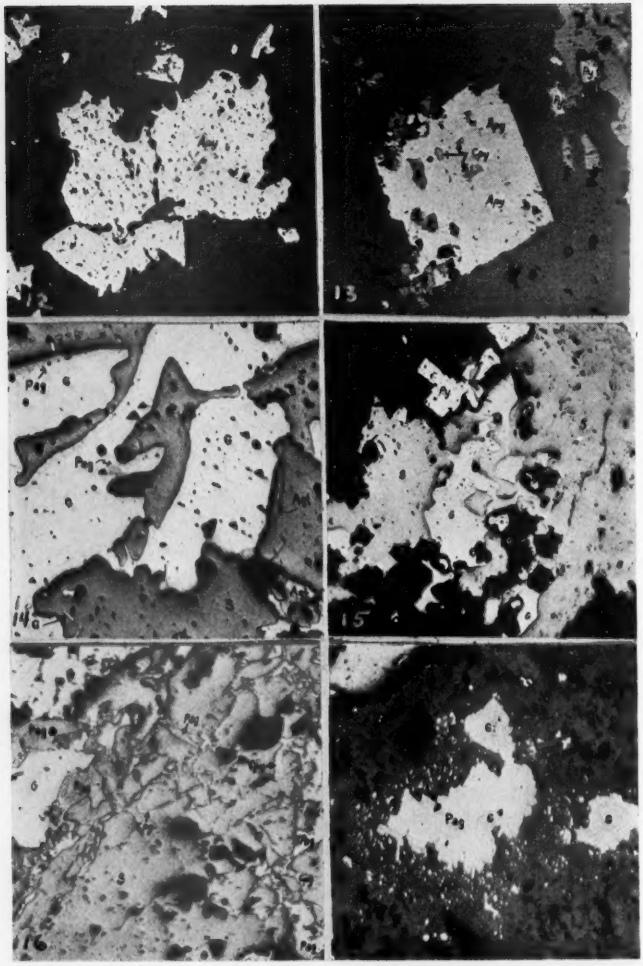


FIG. 12. Arsenopyrite (*Apy*) intergrown with quartz (*Q*). ×90.

FIG. 13. Automorphic crystal of arsenopyrite (*Apy*) containing inclusions of sphalerite (*S*) and chalcopyrite (*Cpy*). Arsenopyrite enclosed in quartz (*Q*). ×90.

simple filling of fractures in this mineral. The interpretation is that the rhodochrosite, while still in the process of crystallization, was fractured; that these fractures were filled by solutions carrying rhodochrosite, quartz and minor amounts of sulphides; that the crystallization of the rhodochrosite continued with the development of additions to old crystals and the formation of free automorphic crystals; that the crystallization of the rhodochrosite proceeded essentially to completion before quartz crystals commenced forming; and that quartz carrying inclusions of sulphides and rhodochrosite formed interstitially within these fractures, assuming in all cases the automorphic outlines of rhodochrosite. Forces which caused the fracturing may have arisen from several sources, but the close association with mineralization processes suggests that the principal forces involved were those incident to the penetration of the ore-bearing solutions.

Sericite in finely divided flakes is associated in small amounts with the earlier, clear quartz. In the later, usually darker colored quartz, this mineral is not ordinarily present. Its abundance with respect to the early formed quartz and marked decrease with the appearance of the early sulphides, suggests that it belongs to the early stage of quartz crystallization. Sericite probably formed contemporaneously with the first quartz and slightly overlaps the earliest sulphides.

Quartz is the most important introduced gangue mineral in the Vipont ores, formed by replacement, chiefly of calcite stringers within the fractured limestone. As a rule, the last formed quartz in the central regions of the orebodies is finer grained than the earlier quartz and contains dark inclusions resembling

FIG. 14. Galena (*G*) interstitial in sphalerite (*S*) and intergrown with pyrargyrite (*Pag*). Note inclusions of galena in sphalerite. ×80.

FIG. 15. Galena (*G*) intergrown with pyargyrite (*Pag*) and interstitial in sphalerite (*S*). Quartz (*Q*). Pyrite (*Py*). ×75.

FIG. 16. Galena (*G*), chalcopyrite (*Cpy*), and pyrargyrite (*Pag*) filling fractures in sphalerite (S). Pyrargyrite intergrown with galena and chalcopyrite. ×250.

FIG. 17. Pyrargyrite (*Pag*) intergrown with galena (*G*) which has replaced calcite (*C*). ×90.

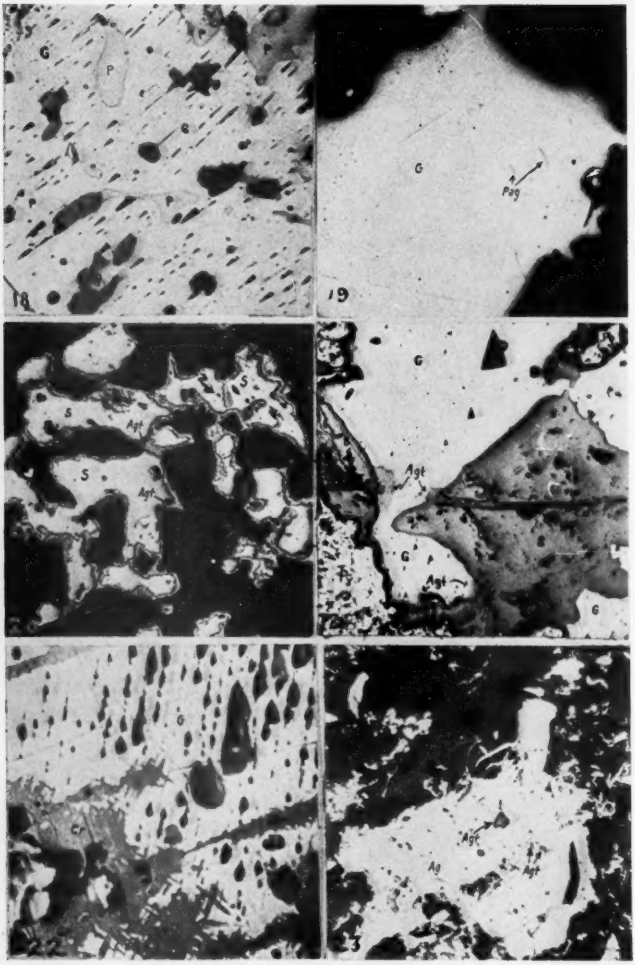


FIG. 18. Inclusions of pearceite (*P*) in galena (*G*). Dark gray mineral is quartz (*Q*). ×90.

FIG. 19. Inclusions of pyrargyrite (*Pag*) in galena (*G*). ×250.

manganese dioxide. It is possible, as has been suggested by Vhay,[12] that this dark material is inherited from the replacement of surrounding limestone which contains similar material. However, this is not in harmony with the fact that the earlier quartz, usually directly associated with the limestone, is clear. As pointed out above, the two quartz types appear to represent phases of the same period of continuous crystallization, since the variation in color and amount of inclusionary material between the two is gradational.

Quartz displays mutual crystalline boundaries with pyrite, arsenopyrite, and sphalerite (Fig. 7), though it ordinarily conforms to the automorphic outlines of these minerals. It has been observed associated with sphalerite as inclusions in crystals of arsenopyrite (Fig. 8), and filling fractures in pyrite, arsenopyrite, and sphalerite (Figs. 9 and 10). Late sulphides usually have an interstitial relationship to the quartz.

Pyrite, constituting the most abundant and prominent sulphide of the ores is almost invariably associated with quartz, arsenopyrite, and sphalerite. Both massive and crystalline forms are known. Large bodies of massive pyrite were reported in the early workings of the mine, where, together with arsenopyrite, it comprised the principal part of the large replacement sulphide bodies just below the Wardlaw shale. Most commly it occurs in the form of cubes and retains its automorphic outline against arsenopyrite (Fig. 11), though some mutual crystalline bounda-

[12] Vhay, *op. cit.*

FIG. 20. Sphalerite (*S*) replaced peripherally by argentite (*Agt*). × 90.

FIG. 21. Galena (*G*) interstitial in sphalerite (*S*). Argentite (*Agt*) replacing sphalerite along fracture. Note that fracture does not extend across galena. Chalcopyrite inclusions in sphalerite at (*X*). Axial mineral in argentite veinlet unidentified. × 90.

FIG. 22. Cerussite (*Cr*) replacing galena (*G*) along cleavage directions. × 90.

FIG. 23. Native silver (*Ag*) containing remnants of argentite (*Agt*). Dark gray mineral quartz (*Q*). × 80.

ries occur between these two minerals. This relationship indicates that, in general, pyrite is older than arsenopyrite, but that there has been some overlap in their crystallization.

Sphalerite commonly conforms to the automorphic outlines of pyrite crystals. Inclusions of pyrite in sphalerite are rather common and show an inclination toward rounding; inclusions of sphalerite are also found in pyrite, but not as often. Deposition of sphalerite was evidently later than pyrite, though a slight overlap is apparent between the two minerals.

Against galena and all the sulpho-salts of silver, pyrite usually retains its automorphic form. These minerals transect pyrite in small veinlets filling fractures and also have an interstitial relationship to quartz that cuts the pyrite in veinlets (Fig. 11). Pyrite, therefore, is considered to have preceded the deposition of galena and all the sulpho-salts of silver.

Arsenopyrite, next to pyrite, is the most conspicuous and abundant of the sulphides comprising the Vipont ores. It has been less susceptible to oxidization than pyrite, resulting in more general distribution throughout the mine, and samples of unaltered arsenopyrite are obtainable from the highest workings. Like pyrite, it occurs in both massive and disseminated crystalline form, the latter predominating. Well formed crystals of arsenopyrite are commonly observed in hand specimens.

In the polished specimen, arsenopyrite appears as a pale yellow mineral, somewhat lighter in color than pyrite. It is nearly always associated with quartz, pyrite, and sphalerite. Mutual crystalline boundaries with quartz are common (Figs. 12 and 13), though automorphic outlines of arsenopyrite against quartz are more prevalent. Late quartz fills fractures in arsenopyrite. Inclusions of chalcopyrite, quartz, and sphalerite in arsenopyrite have been noted frequently (Figs. 8 and 13). Sphalerite commonly conforms to the automorphic outline of arsenopyrite, though mutual inclusions and mutual boundaries are not uncommon. Arsenopyrite is considered to be somewhat older than sphalerite, but overlapping the period of crystallization of that mineral. Galena and the late silver salts retain the same relation-

ships toward arsenopyrite as toward pyrite; they are definitely later.

Gold in the Vipont ores is known only from assays with widely divergent values of this metal. Gold has never been recognized in the hand specimen of these ores, and it is only doubtfully recognized in polished sections. A mineral resembling gold occurs as inclusions, associated with inclusions of chalcopyrite, in arsenopyrite. It is lighter in color than the chalcopyrite and etches black with potassium cyanide. The inclusions are roughly equidimensional, and occasionally exhibit straight contacts with arsenopyrite. The exact age relationship of these inclusions to arsenopyrite is not known, but they appear to be roughly contemporaneous with it and correspond to the same period of crystallization as that to which the chalcopyrite inclusions belong. The samples in which these inclusions were recognized came from lower A–73 raise. The gold-bearing arsenopyrite displays relations with other minerals and other arsenopyrite which suggest that it was among the earliest of the arsenopyrite to form.

Sphalerite is found in unoxidized or only partially altered ore. It is not a conspicuous mineral in the hand-specimen, occurring in smaller amounts than either pyrite or arsenopyrite. It is, however, a rather conspicuous mineral in polished sections, where its recognition is made easy by the fact that it invariably contains abundant inclusions of chalcopyrite. Its color is a pinkish-gray by reflected light. As shown above, sphalerite, though overlapping both arsenopyrite and pyrite, is largely later than both. Automorphic boundaries with quartz are common, but the reverse relation also occurs. Later quartz, however, fills fractures in sphalerite. No sphalerite formed subsequent to this period of fracturing. Galena conforms to the automorphic outlines of sphalerite and is usually interstitial within it (Figs. 14 and 15).

Chalcopyrite is a relatively minor component of these ores and is rarely observed in the hand specimen. As in P–41 raise, however, it has been prominent in some stopes. Ordinarily, it is closely associated with ores carrying abundant sphalerite, pyrite, and arsenopyrite.

Two forms have been observed in polished section: inclusions in sphalerite and arsenopyrite (Figs. 13 and 15), and small veinlets and interstitial bodies in the earlier sulphides (Fig. 16). Inclusions in arsenopyrite are larger, but less abundant than in sphalerite, averaging about 0.03 mm. in the arsenopyrite as compared with 0.004 mm. for the sphalerite. Though rounded contacts with the arsenopyrite are most prevalent, straight contacts, which may represent crystal faces are not uncommon. Inclusions within automorphic crystals of arsenopyrite suggest that these straight contacts are not related to crystallographic directions within the arsenopyrite, but that they probably represent crystal faces of the chalcopyrite.

Inclusions of chalcopyrite in sphalerite are of especial interest. Their invariable presence in the sphalerite of these ores, serves as a criterion in identifying this mineral. In form, the inclusions are distinctly irregular, no one shape predominating. There is, however, some tendency toward elongate grains, the long dimensions of which are parallel to crystallographic directions within the sphalerite. Extension of elongate blebs to the point of coalescence with other blebs in line, has produced vein-like forms within the sphalerite. Rarely these vein-like forms have been observed to intersect, and no apparent thickening occurs at their junction. In most sphalerite grains the inclusions have a bunched distribution, large areas within the grains being apparently devoid of inclusions, while adjacent areas are densely populated. There is a marked tendency for thickly populated areas to occur near but not at the edges of the sphalerite grains. With contiguous grains, such a condition results in the formation of "runs," in which there are no inclusions of chalcopyrite. It is not uncommon for chalcopyrite and galena to be found as elongate inclusions occupying the axial positions of these "runs."

In the polished sections of ores taken from lower A–73 raise, chalcopyrite fills fractures in sphalerite (Fig. 16). It has also been observed as irregular masses intergrown with galena interstitial in quartz and sphalerite. It is noteworthy that wherever chalcopyrite in veinlets or irregular masses comes in contact with

sphalerite, there is a marked increase in the abundance of chalcopyrite inclusions in the sphalerite. A similar relationship has been described by Buerger [13] for the unmixing of chalcopyrite from sphalerite from the ores of Bingham Canyon, Utah. Here it was observed that when sphalerite was heated in contact with chalcopyrite to temperatures of 350 to 400° C. numerous small blebs of chalcopyrite developed within the sphalerite near the contact with the chalcopyrite. Buerger says:

> This development of minute chalcopyrite blebs in the sphalerite adjacent to large masses of chalcopyrite can only be accounted for by a mixing and unmixing process having taken place. The chalcopyrite dissolved in the adjoining sphalerite to form a solid solution of the two minerals. . . . Since the chalcopyrite in solution was more concentrated near the large mass of that mineral, more blebs have been segregated close to the mass, diminishing in number and finally disappearing farther out in the sphalerite.

This may account for the extra density of chalcopyrite inclusions in sphalerite next to chalcopyrite in the Vipont ores. Though it is clear that the larger masses of chalcopyrite are later than the sphalerite, sufficient heat in the mineralizing solutions may have been present subsequent to the crystallization of the chalcopyrite to cause a diffusion of chalcopyrite into solid solution with sphalerite. Later unmixing would thus reveal a concentration of chalcopyrite in sphalerite adjacent to larger chalcopyrite areas. As will be seen, chalcopyrite crystallization was essentially complete before the precipitation of galena and the sulpho-salts of silver.

Galena is a minor constituent of the Vipont ores. It is not commonly recognized in the hand specimen and where observed is usually in small grains. The largest specimens observed came from lower P–2 raise. In nearly all cases galena is found associated with enriched or partially oxidized ore and is usually partially altered to secondary minerals.

In the polished section galena appears as a white mineral usually intergrown with the sulpho-salts of silver interstitially in the

[13] Buerger, N. W.: The unmixing of chalcopyrite from sphalerite: Am. Min., XIX: 528, 529, 1934.

earlier sulphides (Fig. 15). It conforms to the automorphic outlines of quartz, pyrite, arsenopyrite and sphalerite (Fig. 20) and fills fractures in the last three of these minerals. Inclusions of galena in sphalerite are common, though not nearly as abundant as chalcopyrite. These ordinarily have rounded outlines but galena inclusions conforming to the crystal outlines of sphalerite have been repeatedly observed. It is not uncommon for the galena inclusions to occupy "runs" parallel to crystallographic directions within the sphalerite, in which case, the inclusions are linearly arranged elongate, rounded masses, similar in shape, though generally larger, than chalcopyrite inclusions occurring under similar conditions. Galena is considered to be definitely later than sphalerite and to occupy spaces between sphalerite crystals. Small masses of chalcopyrite are intergrown with galena within sphalerite. Galena is also definitely later than pyrite and arsenopyrite and formed after a period of local fracturing and concomitant quartz-deposition.

Galena is commonly associated with highly silicious ores, but in some samples taken from N–43 raise, galena and pyrargyrite directly replace calcite in the wall-rock (Fig. 17). In these same sections, galena, where associated with quartz, always assumes an interstitial relationship with that mineral. Automorphic inclusions of quartz in galena are common, but inclusions of galena in quartz are rare, if not wholly lacking. These relationships indicate that the precipitation of galena and the sulphosalts of silver persisted longer than the deposition of late quartz; locally they replaced the calcite of the wall-rock directly.

In polished sections of ore coming from X–9 stope a mineral has been recognized which, in repeated tests, gave reactions corresponding to those noted by Short[14] for argyrodite, although the presence of germanium has not been verified. It is light gray and possesses slight anisotropism. The mineral occurs intergrown with chalcopyrite interstitially in contact with quartz and

[14] Short, M. N.: Microscopic determination of the ore minerals. U. S. Geol. Surv. Bull. 825: 85, 1931.

pyrite and is considered to be about the same age as the chalcopyrite.

The ruby silver pyargyrite is quite often observed in the ores of the Vipont mine, but is ordinarily too finely disseminated to be definitely identified in the hand specimen. It is almost always associated with galena-rich ore and is usually interspersed in that mineral.

In polished section, this pyrargyrite is light blue and shows a deep red internal reflection. Identification by etch tests was supplemented by microchemical tests proving the antimony content of the mineral, thus distinguishing it from proustite, the arsenical ruby silver. Commonly, pyrargyrite occurs as automorphic inclusions in galena (Fig. 19), though it also occurs intergrown with this mineral. Specimens taken from X–9 stope show a particularly good development of these inclusions. They are irregular, though there is a definite tendency toward elongate-rectangular and diamond-shaped outlines. The average maximum diameter for samples from X–9 stope is about 0.025 mm. The width of longer rectangular forms is usually about one fifth of the maximum diameter. Boundaries are straight and are definitely crystal faces. The orientation and distribution of the inclusions with respect to each other and with respect to crystallographic directions within the galena indicate that no influence was exercised by the galena in their distribution.

Of the primary silver minerals found in the Vipont ores, pyrargyrite is by far the most abundant and is responsible for most of their silver content.

Pearceite, recognizable only in polished section, is almost invariably associated with pyrargyrite. It is slightly more gray than galena, shows slight anisotropism, and has a dark red internal reflection. Its identification is based both on etch and microchemical tests. Wherever it has been observed in contact with the earlier sulphides, the automorphic outlines of these minerals persist. Like pyrargyrite, it is found as inclusions in and intergrown with galena (Fig. 18). Together with pyrargyrite,

it fills fractures in the earlier sulphides. It is considered to be essentially the same age as pyrargyrite and galena.

Tennantite was identified in polished sections of specimens from A–103 raise. It is light gray, about the same color as pearceite, but with slightly higher relief. It occurs intergrown with pyrargyrite and galena and is interstitial to quartz and the early sulphides. It is believed to be essentially contemporaneous with pyrargyrite.

In polished section, a pink mineral in grains too small to be identified is associated with pearceite and tennantite. It appears to be about the same age as the pearceite.

Supergene Minerals. The hypogene ores of the Vipont mine have been extensively modified by supergene processes. The extent to which alteration in the primary ores has proceeded has been dependent on the ease and readiness of access to the ores by the modifying agents. Thus, it is clearly evident that secondary alterations are intimately related to structural features within the orebodies and wallrock. It is not possible to divide the Vipont mine into zones in which either primary, enriched, or oxidized ores predominate. Oxidized and enriched ores are not uncommon in the lowest workings of the mine, and essentially primary ores are not unknown in the highest levels of the mine. It is true, however, that oxidized ores are more abundant in the upper levels, and that supergene enrichment has been greater in the intermediate levels. Single orebodies are usually modified by both enrichment and oxidation. Active groundwater circulation is usually confined to local zones of fracturing and the modification of an orebody depends on its proximity to one of these zones. If a general groundwater table exists in this mine similar to that ordinarily assumed to conform with superficial topographic features, it has not played a part in the division of the ores into zones.

The following minerals have been noted as alteration products of the primary ores resulting from processes of downward enrichment: undetermined lead silver salt, intermediate between galena and argentite; argentite, covellite, chalcocite, cerussite, and possibly native silver. Predecessors of these minerals are mainly

sphalerite, galena, and the sulpho-salts of silver. Arsenopyrite has locally been replaced by argentite.

Occurring between galena and argentite in the enriched ores is a narrow band of a mineral whose formation invariably precedes the replacement of galena by argentite and appears to represent a transition from one mineral to the other. In color and hardness it is like galena and reacts similarly to tests for that mineral.[15] In fresh untreated sections it is indistinguishable from galena, but is brought out in striking contrast by staining the galena with hydrogen peroxide. This relationship is particularly well shown in samples from A–10 raise and lower A–73. Galena has been slightly replaced by covellite, cerussite, and chalcocite; argentite extensively replaces sphalerite; but this intermediate mineral occurs only between argentite and galena. Similar relationships between galena and argentite have been noted by Bastin [16] in the ores of Tonopah, Nevada, and he interprets the transitional mineral as a double sulphide of lead and silver.

Argentite is by far the most abundant silver mineral of the enriched ores and is economically the most important. Ordinarily it is not recognized in hand specimen, though locally it has been found in fairly large masses, intimately associated with wire silver. Argentite is found in nearly all polished sections of the sulphide ore, regardless of the part of the mine from which the specimens come. It is moderately light gray with a marked green tinge. It replaces galena and sphalerite (Figs. 20 and 21) and, locally, arsenopyrite. Alteration of the ruby silvers has mainly been to argentite. Sphalerite displays striking peripheral replacement by argentite, though in places its crystallographic directions clearly control the replacement. Laths of covellite, intimately intergrown with the argentite, are common in the replacement of both sphalerite and galena, and a small amount of chalcocite is often associated with the covellite. Small veinlets of argentite in sphalerite sometimes have a central parting of an

[15] Short: *ibid.*, p. 73.

[16] Bastin, E. S.: The genesis of the ores at Tonopah, Nevada. U. S. Geol. Surv. Prof. Pap. 104: 21–23, 1918.

apparently earlier mineral which may be cerussite or, quite possibly, a silver haloid. This mineral has not been observed to replace sphalerite directly, but is always separated from it by argentite. It may represent a replacement of elongate galena inclusions which formerly occupied the spaces between intergrown sphalerite crystals, since galena with such relationship to sphalerite is quite common in the primary ores. Replacement of galena by argentite has not been as extensive as the replacement of sphalerite.

Corresponding to the observations of Foss,[17] the present writer has observed argentite "between quartz grains . . . and filling minute fractures in these same grains." However, in such cases, the argentite is almost invariably intergrown with small amounts of covellite or is connected with argentite which, due to its relations and associations with other minerals, can definitely be considered to be of secondary origin. Evidence which would conclusively indicate a primary origin for argentite in these ores has not been observed.

Covellite and chalcocite, rarely recognized in hand specimens, occur only in small amounts as replacements of chalcopyrite, sphalerite, and galena. Both minerals commonly occur intergrown with argentite in the replacement of galena and sphalerite. As seen in samples from A–73 raise, covellite, intimately intergrown with cerussite, is found replacing galena.

Cerussite often replaces galena (Fig. 22), and is occasionally seen as a coating on ores rich in galena.

A mineral which resembles native silver in etch tests and color is often found intergrown with argentite in enriched ores. Inclusions of this mineral in argentite are too small to make determinative tests.

Oxidized Minerals. A large number of minerals have been recognized in the oxidized ores, but it was generally impossible to discover any consistent relationships between them. They are native silver, malachite, azurite, cuprite, native copper, orpiment, realgar, cerargyrite, limonite, hematite, anglesite, and calcite.

[17] Foss: *op. cit.*, p. 18.

Many of these minerals occur in very small quantities and not always in association with other minerals of the group. It is therefore impractical to attempt to relate them to any particular phase of the oxidization processes.

Next to argentite, native silver is probably the most important mineral, occurring abundantly in locally oxidized and enriched ores throughout the mine, but with the largest amounts in the intermediate levels.

Native silver occurs in the ores most commonly as wire, in thin flakes, and in well-formed, tiny crystals. Druses in the richer ores are often partially filled with curling wires of native silver, generally associated with minor amounts of argentite from which the silver probably emanated. Wire silver has been observed up to one eighth of an inch in diameter, but ordinarily it is smaller. On larger wires of silver, there are developed nicely formed tiny cubic crystals of silver which give the wire a punctate appearance. Tiny flakes of silver have been observed clinging to the limestone hanging wall of the orebodies, or plastered on masses of sulphide. Native silver frequently replaces argentite (Fig. 23).

Malachite and azurite are locally present as stains and as fracture-fillings throughout the oxidized ores.

Cuprite was recognized in association with a small amount of native copper in a specimen of partially oxidized ore from K-level. Small amounts of pyrite and arsenopyrite still remained in the specimen, but none of the earlier copper minerals could be recognized.

Orpiment and realgar occur in small amounts as thin, dust-like coatings on arsenopyrite in partially oxidized ores.

Cerargyrite may occur in the ores. A small amount of material taken from oxidized ores on K-level has the general appearance of cerargyrite and reacted positively to a microchemical test for silver. That the mineral is probably present in the ores is suggested by the presence of silver in a solution derived from treating the oxidized ore with concentrated ammonium hydroxide.

Limonite and hematite occur in all the oxidized ore and are

locally prominent throughout the mine. Hematite imparts a much deeper color to the ores than limonite.

Secondary calcite is prevalent, and locally, cavities are lined with large well formed crystals.

Anglesite is found associated with cerussite locally in oxidized zones, but never occurs in large quantities.

Though sphalerite is an important component of the primary ores, no secondary zinc minerals were found. That they are present, however, is suggested by small quantities of zinc in assays of the oxide ores. It is likely, considering the country rock and the usual presence of considerable calcite in the oxidized ores, that smithsonite is the form in which the secondary zinc exists.

Paragenesis. Due to two episodes of fracturing, the deposition of the hypogene minerals can be divided into three periods. The deposition of rhodochrosite is separated from the deposition of the later minerals by a period of fracturing, the fractures being filled with quartz and minor amounts of sulphides. Galena and the sulpho-salts of silver are interstitial in quartz that fills fractures in pyrite, arsenopyrite, and sphalerite. The three phases of the mineralization thus indicated are (1) a period in which rhodochrosite was formed to the exclusion of other minerals, (2) a period in which the dominant minerals formed included quartz, pyrite, arsenopyrite, sphalerite, and minor amounts of chalcopyrite, and (3) a period in which galena and the sulpho-salts of silver predominated over all other minerals. Economically, the last of these periods has been by far the most important, though the second phase appears to have produced the gold values of the ore.

The paragenetic sequence of the ore minerals is shown grahpically in Table 2.

Tenor of the Ores. In discussing the tenor of the ores it is necessary to distinguish between the kinds of mining methods used in recovering the ore. It is evident from the patchy distribution of the ore, and the variation in size and form of the orebodies, that the average value of recovered ore depends directly on the method of recovery. Operations in quest of ore suitable

TABLE 2.

PARAGENETIC SEQUENCE OF THE MINERALS IN THE VIPONT ORES.

Minerals.	Hypogene Minerals.	Products of Supergene Enrichment.	Products of Oxidation.
Rhodochrosite	—		
fracturing			
Sericite	—		
Quartz	—		
Pyrite	—		
Arsenopyrite	—		
Gold	—		
Sphalerite	—		
Chalcopyrite	—		
fracturing			
Galena	—		
Argyrodite	—		
Pyrargyrite	—		
Pearceite	—		
Tennantite	—		
Unidentified*	—		
Unidentified†		—	
Argentite		—	
Covellite		—	
Chalcocite		—	
Cerussite		—	x
Native silver		—	x
Malachite			x
Azurite			x
Cuprite			x
Native copper			x
Orpiment			x
Realgar			x
Limonite			x
Hematite			x
Calcite			x

* Unidentified mineral associated with sulpho-salts of silver in primary ores.

† Unidentified mineral intermediate between galena and argentite in the enriched ores.

for milling, and operations in quest of high grade ore suitable for direct shipment have both been carried on. Nearly all the high grade smelting ore recovered is screened, much of it underground. The values are principally found in the minus ½ inch fines.

Two hundred composite analyses, each composed of ten grab-samples from freshly blasted high grade ore at different localities throughout the mine, show an average silver content of 27.06 fine ounces, and an average gold content of 0.078 ounce per ton. Individual orebodies have been known to carry as much as 10,-

000 ounces of silver and 4 to 5 ounces of gold per ton. The ratio of gold to silver is highly variable and there seems to be no consistent relationship of tenor to locality. Out of 63,710 tons of ore mined during 1922, the average recovery per ton was 0.046 ounce of gold and 16.3 ounces of silver.

Temperature of Formation of Primary Ores. Deposition of the minerals of the primary sequence is divisible into three general phases. All of the minerals of the first two phases are found under a wide range of temperature conditions and are thus not useful in the determination of the temperature at which the ores were formed. The sulpho-salts of silver in the later phase, on the other hand, are characteristically developed under low temperature conditions. Since chalcopyrite overlaps the deposition of these minerals and the earlier sulphides, it appears that deposition was from cooling solutions which even in their first stages had temperatures not much above those at which the sulpho-salts of silver were formed. The fine-grained texture of the ores corroborates the conclusion that the ores were formed under low temperature or epithermal conditions.

Character of Hypogene Solutions. As carbonates are in general deposited only from neutral or basic solutions, the appearance of rhodochrosite as the initial mineral of the primary sequence would seem to indicate that the early mineralizing solutions were of this character. The appearance of silica has no particular significance in the interpretation of the character of the solutions; this applies also to the presence of arsenopyrite, chalcopyrite, galena, and the sulpho-salts of silver. The presence of sphalerite suggests neutral or basic solutions, though this mineral may be precipitated from either acid or basic solutions. Wurtzite, which would indicate acidity, has not been recognized in these ores.

The presence of pyrite rather than marcasite also suggests basic or neutral solutions. Dana [18] states that when pyrite is deposited from rising solutions, these solutions are usually basic.

A final bit of evidence as to the composition of the solutions

[18] Dana, E. S.: A Textbook of Mineralogy, 4th edit. rev. John Wiley & Sons, Inc., New York, 1932, p. 434.

lies in the type of wall-rock in which the mineral deposits were formed, that is, it is unlikely that acid solutions could exist long in contact with the carbonates of the wall-rock.

It is concluded that at least during the main period of the mineralization the ore-forming solutions were either neutral or slightly alkaline, more likely the latter.

Other Mines and Prospects.

Skoro Mine. This mine is situated approximately a thousand feet southeast of the Vipont camp and has about the same elevation as A-portal of the Vipont mine. It was originally located in 1922 by a Mr. Skoro, then employed by the Vipont Silver Mining Company, and is opened by a 763 foot tunnel through the lower porphyry into the underlying Sentinell limestone, from which drifts have been driven in three directions. In all, approximately 1,300 feet of development work have been completed.

No important shipments of ore have been made from this mine. The tenor of the ore has been very erratic and, except for small pockets, has been generally low. The ore shipped has all been carefully selected screenings of the mined ore. Like the Vipont mine, the values are principally found in the fines. During the process of development work in the fall of 1936 the following assays were run.

SOME ASSAYS OF THE ORES FROM THE SKORO MINE.

Assay.	Gold.	Silver.	Location.
	Ounces per Ton.	Ounces per Ton.	
1	0.00	1.68	Unknown
2	0.34	152.12	"
3	0.06	14.04	East drift
4	0.005	2.80	North drift
5	tr.	2.00	" "
6	tr.	1.68	" "
7	0.02	10.70	" "
8	0.015	5.50	" "
9	0.02	19.37	" "

The occurrence is similar in many respects to that of the Vipont mine, though all the ore thus far found is confined to the

Sentinell limestone. The ore occurs as thin, scattered stringers intimately associated with stringers of secondary calcite along the bedding planes of the limestone. Rhodochrosite is locally abundant and, as in the Vipont mine, extensively replaces the secondary calcite stringers. Silicification of the wall-rock has not been as extensive as in the Vipont mine. The sulphide ores are usually soft and are not readily adaptable to study in polished section. A close examination of hand specimens, however, revealed essentially the same group of primary minerals present in the Vipont ores, though these ores are not nearly as extensively altered by processes of oxidation and enrichment.

Sentinell Group. The Sentinell workings are a group of shallow workings in the Sentinell limestone, situated about 1,500 feet northeast of A-portal of the Vipont mine (Fig. 1). They are very irregular, and the ore seems to have been found in patches and pockets along small slips and crenulations in the limestone. Except for some prospecting, active work has not been done in this group of workings since about 1922, when a small shipment of high grade ore was made from the lower workings. It is noteworthy that, compared with the Vipont ores, these ores have higher gold values. The ore is highly oxidized, though small pieces of essentially unaltered sulphides have been found.

Francis-X Mine. This mine was caved at the time of the writer's visits to the district and the workings were not accessible. The mine consists of a tunnel at equal elevation with the A-portal of the Vipont mine driven S. 75° E. through the basal part of the Phelan limestone into the Wardlaw shale. So far as known, the base of the shale has never been reached and no ore has been removed from the mine.

Midway Tunnel. Accurate information about this tunnel was not available. The tunnel is approximately 1,550 feet long and is driven through the Wardlaw shale into the Vipont limestone. The elevation of the portal, equal to that of the Phelan tunnel, is 7,201 feet above sea level. When active exploratory work by the Vipont Silver Mining Company ceased in 1923, A-level of the Vipont mine was being driven westward to connect with a raise

to be driven from the Midway tunnel. Accurate information as to the occurrence of ore in the Midway tunnel was not available to the writer, though Mr. H. K. Lake states that some was found in an east drift. Before this could be removed the tunnel caved and has not been opened since.

Lexington-Argenta Workings. These workings are the site of the first discovery of ore in the district. They are located from 100 to 500 feet below the Midway tunnel and 2,000 feet to the west. The Argenta workings consist of three tunnels driven along a vertical break in the Phelan limestone. At present, they are largely inaccessible. A little stoping appears to have been done in the lowest of the tunnels and a small amount of ore may have been extracted. The ore appears to be in the form of small pockets in the limestone. These mines have not been worked for more than 30 years.

The Lexington group, on the south side of Lexington Canyon, consists of a few short tunnels driven southward through the Phelan limestone to the contact with the lower porophyry. These tunnels are all caved at the present time and information as to the nature of the underground workings was not available. Ore which could certainly be considered to have come from these workings was not found.

Dolly Clark Mine. The Dolly Clark mine is the westernmost mine in the district. It consists of a tunnel 600 feet long driven just below the contact of the lower porphyry with the Phelan limestone into the Wardlaw shale. Several short drifts and winzes have been driven off the main tunnel in quest of ore, but up to the time of the writer's last visit to the mine, no appreciable values had been discovered. The mine has been shut down since 1937.

ACKNOWLEDGMENTS.

In the preparation of this paper, the writer gratefully acknowledges the help and cooperation of his friends who are now or were formerly at Vipont. Especially does he express his sincere thanks and appreciation to Messers Ridgway Spencer, Frank N. Spencer,

and Leo Duquett for their willing cooperation and the wealth of information which each so kindly volunteered. Thanks are also due the property owners of the district for the privilege of studying their respective claims, most notably the Vipont Silver Mining Company for free access to the Vipont mine and maps and information pertaining thereto. During the course of his laboratory work, it was the writer's privilege to work under the guidance of Dr. E. S. Bastin of the University of Chicago. For the many helpful suggestions he so kindly offered, the writer expresses his sincere appreciation.

MT. VERNON, ILLINOIS,
May 11, 1942.

JOINTING IN THE COAL BEDS OF OHIO.

KARL VER STEEG.

ABSTRACT.

Data obtained by field work and from engineers and operators, have revealed some important facts about jointing in the coal beds of Ohio. The joints show remarkable uniformity in trend, and the direction of the cleat joints appears to be the same even though more than one bed of coal is involved. In the northern part of the coal mining area, the joints occur usually in two sets known as the face and the butt joints which lie at right angles to each other. One set extends in a northeast-southwest direction and the other has a northwest-southeast trend. Farther south, one set extends in a direction a few degrees west of north and the other set, at right angles, has a trend a few degrees north of east, or nearly east and west. The joints are arranged in an arc, the convex side to the west. This curve corresponds to the curve in the folded Appalachians in central and eastern Pennsylvania, in the same latitude.

There is varied opinion as to the origin of the cleat in coal, but it is difficult to explain the remarkable uniformity in direction of the joints, the fact that in successive coal beds in the same area the cleat joints have approximately the same trend, and the parallelism with the folded Appalachians to the east in the same latitude, unless one believes in the theory of tectonic origin. The joints were probably formed by deformation at the time the folded Appalachians were formed, although one must admit that shrinkage has taken place to some extent and that this fact cannot be denied.

ALL users of bituminous coal are familiar with the fact that it has a marked tendency to break in blocks, and that this phenomenon is a more distinct feature of some coals than of others. Anthracite and cannel coal break with conchoidal fracture, and other bituminous coals generally break with a rectangular or cubical fracture. The vertical joint planes of coal beds or "cleat" are smooth, often lustrous planes, which run at right angles or diagonally to the bedding plane. They may cause the very desirable cubic or rectangular breakage of many coals. The cleat is closely

spaced, and as a rule, possesses a fairly uniform strike. Moore[1] states, "In coal beds there are usually two sets of joints resulting from the drying out of the rocks and the movement of the strata and these run approximately normal to each other. Those which lie normal to the strike and cut across the bedding of the coal are frequently known as cleats. They are, as a rule, more clearly marked than the joints running in the other direction."

Cleat is much better developed in some coals than in others. According to Price,[2] state geologist of West Virginia, "the cleat in the Pittsburgh, Redstone and Sewickley coals is especially well developed. In the Pocahontas and other low volatile coals, while cleat is present, it is harder to measure. The best place to measure the cleat is in the mine on the joint faces where the coal is not disturbed. We find, as a rule, that the fracture pattern is quite consistent and remains regular over a rather large area."

Mine operators take advantage of the cleat for more economical mining. They lay out or develop their mines in conformity with the joint pattern, which makes it easy to determine the direction of the joints by the mine map lay out. The workings of coal mines are best oriented so that the operations are mainly directed at right angles to the cleat in order to obtain the largest falls of coal. If coal beds are mined at right angles to the cleat, great masses of coal may burst suddenly into the mine room. In the Ohio mines, as well as in other localities, the coal is mined by driving entries at right angles to the face joints, which are the best developed. The butt entries lie at right angles to the butt joints which usually occur at approximately right angles to the face joints. From the direction of the entries the trend of the joint systems can usually be obtained.

It has been observed that the joints become wider and more distinct in the coal bed if the workings are idle for some time. Minerals, such as calcite, iron pyrite, gypsum, limonite, marcasite and others may occur in the cleat joints.

[1] Moore, E. S.: Coal, Its Properties, Analysis, Classification, Geology, Extraction, Uses and Distribution. John Wiley & Sons, New York, 1922. P. 8.

[2] Price, P. H.: Personal Communication.

Stutzer and Noe [3] say that cleat does not continue across shale partings which may divide the coal bed into two or more benches, nor do the cleat joints usually continue into the roof or into the fire clay. The jointing in the roof and fire clay may have a different orientation. In England the strike of the cleat in a succession of parallel coal beds is said, usually, to be the same. They quote various observations on cleat, as follows:

According to Frieser [4] (1914), the production of cleat by lateral pressure is plainly apparent in the Falkenau basin of northern Bohemia. Not only are the cleat joints in the coal bed parallel with each other, but they are also parallel with the other fissures in the region. They also have a northwestern or Hercynian direction, indicating that the pressure came from the southeast. The Agnes coal bed was fractured by this pressure. The cleat surfaces in this coal are not always perfect planes, being characteristically undulating. Even today the seam is under great tension and cracking noises are heard in the workings and suddenly coal walls are thrown down.

Kaiser [5] (1908) described the cleat in the lower Rhenish brown coal district. A few cleat joints are very distinct and usually have smooth walls. Horizontal tree trunks which lie across the cleat are crossed by smooth joint planes. Faulting is usually not associated with cleat. With respect to the origin of cleat, it is significant that it has the same strike for a great distance. Numerous cleat joints run parallel to the fault which cuts the brown coal of the foothills. Kaiser thinks this cleat structure is produced by tectonic processes.

Höfer [6] (1915) believes that cleat must have originated after the strata overlying the coal bed were deposited, since no roof material infiltered into the cleat joints, and according to him the development of cleat is not a tectonic phenomenon. He reaches this conclusion because, although the direction of cleat is gen-

[3] Stutzer, Otto, and Noe, A. C.: Geology of Coal. Univ. of Chicago Press, Chicago, 1940. Pp. 256–257.

[4] *Op. cit.*, p. 257.

[5] *Op. cit.*, pp. 257–258.

[6] *Op. cit.*, p. 258.

erally the same in parallel coal beds, it is not always so. There are instances in which some beds in a series will have cleat running in another direction. He believes that the cause of cleat is inherent in the coal itself and is due to the loss of substance (loss of water, the liberation of carbon dioxide and methane, the rearrangement of the carbon compounds, etc.) which the coal has suffered with its advancement in rank. He therefore considers the cleat joints to be contraction fissures.

Stutzer [7] thinks that cleat joints are parallel rupture planes produced by pressure and does not concur in the explanation offered by Höfer. He believes the uniform strike of the cleat joints in parallel coal beds can be explained only as the effect of pressure.

The above statements indicate that the question as to the origin of jointing in coal beds is not a closed one.

The author has gathered data on jointing in the coal beds of Ohio by observations and measurement in the field, and from information furnished by operators of coal mines, surveyors and mining engineers. Three hundred letters were sent to engineers and operators requesting information concerning the direction of the joints. The results of this survey are indicated on the accompanying map (Fig. 1). Data were obtained from every coal field in Ohio. The beds involved include those seams indicated by the numbers 1–8. The beds most frequently mined and by far the most important from the standpoint of production are the No. 6 (Kittanning) and No. 8 (Pittsburgh). The data resulting from the writer's investigations reveal some important facts.

A glance at the map shows the remarkably uniform trend of the joints which occur in two systems, at right angles or approximately so, and referred to by all engineers as the face and butt joints. The beaded strike lines, plotted on the map, represent the direction of the face joints. In approximately four-fifths of the cases cited, the face joints have a northeast-southwest trend, the number of degrees east of north varying somewhat. One opera-

[7] *Op. cit.*, p. 258.

FIG. 1. Map showing the strike of the joint planes in the coal beds of Eastern Ohio. The face joints are indicated by the beaded lines.

tor reports that in Muskingum county the face joint is a wavy plane and that the course will vary somewhat in different localities and in different coal seams. He reports the butt joints as very irregular. From an operator who is mining the No. 1 (Massillon) coal near North Lawrence, Ohio, comes the observation that the direction of jointing is not the same in all mines

although it does not vary more than 6 to 8 degrees. A mining engineer at New Lexington, Ohio, in the Hocking Valley coal field, reports that the face cleats are not so prominently displayed at one mine as they are at other mines in which he has worked. Another mining engineer from Jackson, Ohio states that the face and butt joints in his locality are present in five or six seams of coal mined in that area, but are not pronounced, and refers to the excellent development of the cleat joints in the No. 8 (Pittsburgh) seam in the vicinity of Pittsburgh. An engineer of wide experience recognizes the fact that face and butt joints are very common in coal seams, but states that he has seen coal with no semblance of these joints. An engineer with much experience in Ohio coal mines reports that the face joints in the No. 6 (Middle Kittanning) seam in Perry and Athens counties are not well defined and their direction varies as much as 5 to 10 degrees from the average. In several other cases the reports indicate that the joints are poorly developed or not clearly defined, or that the joints occur as a wavy plane, their strike varying several degree from the average. However, the majority report two well defined joint systems, and with few exceptions, at right angles to each other. In districts where more than one coal seam is mined, the direction of the joints is the same; the strike of the cleat in a succession of parallel coal seams is the same.

A study of the map shows an unusually uniform trend of the strike. The trend of one set of joints in Portage, Mahoning, Stark and Columbiana counties, in the northern part of the coal mining area, is 25 to 30 degrees east of north. Proceeding southward into Carroll, Tuscarawas, Jefferson, Coshocton, Harrison, and Belmont counties, we find the strike of the same set of joints to be somewhat less than to the north; approximately 10 to 20 degrees east of north. Still farther south, in Muskingum county, the strike of the same set of cleat joints is north and south, and in Perry, Hocking and Athens counties the strike swings to the west of north. If a line were drawn through the strike lines of this set of joints from Mahoning through Athens and Meigs counties, it would make a broad, unsymmetrical curve

or arc, the convex side being to the west. If this curve is compared with the trend of the folded Appalachians in the same latitude in Pennsylvania, to the east, one is impressed by the parallelism. This similarity of strike is not a coincidence. It is probable that the jointing in the coal beds of Ohio is the result of the same tectonic forces which produced the folded Appalachians. If shrinkage and contraction were the sole cause for the production of the cleat joints, one would expect them to be uniform in direction. They should occur in all directions. The remarkable uniformity in direction, the fact that in successive coal beds in the same area the cleat joints have approximately the same strike, and the parallelism with the folded Appalachian mountains to the east in the same latitude, is evidence that the joints are the result of deformation.

College of Wooster,
Wooster, Ohio,
June 20, 1942.

DISCUSSION AND COMMUNICATIONS

GRANITE AND ORE.[1]

Sir: Augustus Locke, in his article, "Granite and Ore," [2] recognizes that many masses of granitic rock are metamorphosed sediments rather than intrusive bodies and raises the question of whether or not the ore deposits associated with them may likewise have been derived from the original sediments. He visualizes the metamorphic granite as simply recrystallized schist and quartzite, and the ore, as derived from the rocks of that particular setting. The formation of the ore, therefore, is entirely independent of any magmatic source.

If it is true that granite may be formed by recrystallization of sedimentary rocks, then it may well be that ore may be formed in the way that Locke suggests; but there are probably few geologists who are likely to admit that granite may be formed by mere recrystallization of sedimentary materials. On the contrary they will probably demand that magmatic emanations be called upon to assist in the transformation of sedimentary rock to granite, particularly to supply the alkalies needed to convert the schist and quartzite to a feldspathic rock. McKinstry, in his discussion of Locke's paper [3] has presented arguments unfavorable to Locke's concept of "metamorphic" granite and ore and advances the general idea of "metamorphic" granite as a product of metasomatic replacement, accomplished through the agency of solutions which themselves came from magmatic sources. He

[1] Published by permission of the Director, Geological Survey, U. S. Department of Interior.

The writer wishes to express his gratitude to G. F. Loughlin, C. P. Ross and E. Callaghan for suggestions made during preparation of the discussion.

[2] ECON. GEOL., 36: 448–454, 1941.

[3] ECON. GEOL., 36: 829–832, 1941.

looks upon this matasomatic alteration as an early, super-intense type of wall-rock alteration and the ore deposits as later, cooler, and commonly satellitic phases of the same process.

For some years the writer has had the opportunity to study "metamorphic" and intrusive granite and the ore deposits supposed to be related to them; and, consequently, has assembled some data that bear on the problem of granite and ore. These data do not support Locke's concept of granite as recrystallized schist and quartzite. On the contrary they support the view that "metamorphic" granite exists only because emanations of magmatic origin have contributed substances necessary to transform the sedimentary rock to granite, the transformation process, as pointed out by McKinstry, being one of metasomatic replacement. Each body of "metamorphic" granite that was observed was located in the marginal zones of batholiths and situated in such a way as to indicate that the granitization of the sedimentary rock was associated with the intrusion of magma, essentially as a phase and product of contact metamorphism. These masses of granitized sediments apparently contain no associated ore deposits, a fact that may suggest that the conditions favorable for forming "metamorphic" granite are unfavorable for forming ore.

Masses of granitized sedimentary rock examined include that of the contact zone of the Cassia batholith in south Idaho.[4] It was pointed out that, had there been no granitization or impregnation of quartzite by "igneous material," the exposed part of the batholith would have had only one third its present size. Exceptionally pure quartzites at that locality have been transformed into granite that is generally porphyritic and in places gneissic. This replacement granite encases the intrusive core, and also forms an essential part of the batholithic mass. The "granitization" of the invaded quartzite has necessitated the addition of potassium and sodium from magmatic sources. In no other way can the abundant deposition of albite and microcline at

[4] Anderson, A. L.: Contact phenomena associated with the Cassia batholith, Idaho. Jour. Geol., 42: 376–392, 1934.

the expense of quartzite be explained; certainly not by recrystallization of the pure quartzite alone. The relation of the "metamorphic" granite to the intrusive granite core and the transition of the batholith outward into quartzite express clearly defined contact phenomena, emanations from the magma playing the dominant rôle in the process. Pegmatites cut the batholithic mass, but there is no closely associated ore.

Parts of the Idaho batholith also have been studied and much "metamorphic" granite has been observed in the marginal zone.[5] Just how much of the Idaho batholith is composed of "granitized" sedimentary rock is not yet known, but evidence is accumulating that certain parts of it were largely sedimentary before being changed by metasomatic replacement to granitic rock. The transformations have not been much different from those expressed in the contact zone of the Cassia batholith. Although there are many ore deposits in and near the Idaho batholith, none of them can be linked with the "metamorphic" granite.

A considerable part of the Nelson batholith in northern Idaho has also been examined and the conclusion has been reached that, had it not been for extensive granitization, the Selkirk Mountains there would have been carved in sedimentary strata instead of the huge mass of "metamorphic" granite that now composes them.[6] Were it not that remnants of original bedding are remarkably well preserved throughout the granitized mass, the rock might easily be mistaken for mechanically intruded gneissoid granite, but the evidence to the contrary is clear. No deposits of ore are known to occur in or near these rocks.

In and about each of the three batholiths, the sedimentary rocks transformed to granite were largely siliceous rocks, in part made up of very pure quartzite. These strata traced inwardly into the batholithic masses gradually lose their characteristic sedimentary features and by progressive increase in feldspar change to an igneous-looking rock that cannot be clearly distinguished

[5] Anderson, A. L., and Hammerand, V.: Contact and endomorphic phenomena associated with a part of the Idaho batholith. Jour. Geol., 48: 561–589, 1940.

[6] Anderson, A. L.: Geology and metalliferous deposits of Kootenai County, Idaho. Idaho Bur. Mines and Geol. Pam. 53: 17–20, 1940.

from magmatic granite. It is as though magmatic emanations of hydrothermal character had soaked ahead of the intruding body of magma and had so "granitized" the invaded rock as to change it into a part of the batholithic body itself. Emanations associated with the closing stages of consolidation also have altered the main intrusive mass [7] as well as the bordering "granitized" rock.[8] Since the completely granitized rock has all the minerals of the magmatic rock, it is not everywhere possible to distinguish between them. It is only in the outer parts of the batholiths that the truly hybrid character of the rock may be revealed.

Although additional quantities of emanations were given off at the close of each batholithic intrusion, forming pegmatites and in some places silicious veins, no orebodies are known to have been made in spite of earlier interpretations to the contrary. The absence of ore deposits associated directly with these batholithic masses suggests that the conditions of differentiation may have been unfavorable for the formation of ore, although the possibility that the temperatures were too high to permit sulphides to deposit within or at their margins or that the emanations were deficient in sulphur cannot be dismissed as reasons for the lack of ore.

Such ore deposits as do occur in or near the batholithic masses appear to be related to intrusive dikes and stocks that are much younger than the batholiths and emplaced under very different conditions. Some of these ore deposits, as in Boise Basin, Idaho, are grouped along Miocene "porphyry belts" that lie within the Idaho batholith.[9] In fact, most of the ore deposits that the writer has examined in Idaho show close genetic associations with intrusive magmas either of early Tertiary or of mid-Tertiary age. Hence, they are wholly independent of the batholiths, which are known to have formed in late Mesozoic time, particularly during

[7] Anderson, A. L.: Endomorphism of the Idaho batholith. Abs. G. S. A., Bull. 51, No. 12, pt. 2: 2016, 1940.

[8] Anderson, A. L., and Hammerand, V.: op. cit.

[9] Anderson, A. L.: Geology and ore deposits of Boise Basin, Idaho. U. S. Geol. Surv., Bull. in preparation.

the Cretaceous period.[10] Many of the deposits now considered to be Tertiary include those formerly assigned to a source in the Idaho batholith. This relation of the ore deposits to Tertiary intrusive rocks is to be discussed in a paper on the classification of ore deposits in Idaho, reserved for future publication.

Some of the ore deposits, however, show very close associations with local granitizing solutions, which, like the ore-forming solutions, followed channels such as those found near magmatic granitic bodies. Examples include the tungsten deposits of the Blue Wing district in Lemhi County, Idaho, and the silver-lead deposits in the Coeur d' Alene district.

In the Blue Wing district the tungsten ores show a marked zonal distribution about an irregular mass of granite exposed in the underground workings at the Ima mine.[11] This granite occurs near the center whereas the ores are in veins along the crest and flanks of a small and considerably fractured anticline of folded quartzitic beds belonging to the Belt series (pre-Cambrian). Intrusion of the magma was accompanied by emanations which in part acted in advance of the magma and in larger part during the closing stages of consolidation. Because of the apparent close structural and mineralogical relations of the veins with the granite and the zoning of minerals away from the contact, the late magmatic emanations seem to play an appropriate rôle in the origin of the deposits and may possibly be looked upon as the earliest phase of the mineralizing process.

The relative small quantity of emanations that spread out in advance of the magma caused some sodic plagioclase and biotite to form in the immediately adjacent quartzite; but the most no-

[10] Ross, C. P.: Mesozoic and Tertiary granitic rocks in Idaho. Jour Geol., 36: 673–693, 1928.

Reed, J. C.: Geology and ore deposits of the Warren Mining district, Idaho County, Idaho. Idaho Bur. Mines and Geol. Pam. 45: 8, 1937.

[11] Anderson, A. L.: Geology and mineralogy of the tungsten deposits in the Blue Wing district, Lemhi County, Idaho. U. S. Geol. Surv., Unpublished manuscript, 1939. This district has since been studied and a report published by Callaghan, E. and Lemmon, D. M.: Tungsten resources of the Blue Wing district, Lemhi County, Idaho. U. S. Geol. Surv. Bull. 931–A: 1–21, 1941.

table changes came during the later closing stages of consolidation when enough potash and silica were added to the intrusive rock and the narrow casing of altered quartzite to convert both to a muscovite granite, which outwardly shows sharp transition into quartzite. These changes are revealed in thin sections which show that the plagioclase in the intrusive igneous rock and in the altered quartzite has been largely replaced by orthoclase and microcline, and that only remnant inclusions of the plagioclase and of quartz grains inherited from the quartzite remain. Much of the biotite of the original rock and in the bordering quartzite has been changed to muscovite and most of the plagioclase that escaped replacement by potash feldspar has been converted to aggregates of coarse-grained sericite. Much lobate quartz penetrates the feldspars. Fluorite, pyrite, and the more ordinary accessory minerals have also been added.

The alteration of the rock suggests that the late emanations were at the start notably rich in potash. While temperatures were relatively high, these emanations deposited potash feldspar, first orthoclase and then microcline (in part by alteration of the orthoclase); as temperatures declined, they deposited muscovite and coarse-grained sericite (by replacement of biotite and remaining plagioclase). As the solutions became less potassic and more silicic, quartz was added to the rock, perhaps about the time the muscovite and sericite were formed. Certain other constituents, such as fluorine, sulphur, iron, and calcium, which had remained in solution, formed fluorite, pyrite, and other accessory minerals. Not all the fluorite and pyrite in the granite may have been formed at this time, however, for much may have been added from the later ore-bearing solutions that rose along fractures in the granite and quartzite.

As will be pointed out directly, the composition of the late magmatic solutions, which altered the granite and locally "granitized" the quartzite, is not greatly different from the composition of the succeeding ore-forming solutions. The early magmatic as well as the later ore-forming solutions contained notable concentrations of potash, silica, fluorine, and sulphur, but

differed in the presence or absence of certain of the metals. However, during the ore-forming stages the temperatures of the solutions were apparently lower than during the late or post-magmatic stage; consequently, the mineral succession during ore deposition began with sericite or muscovite rather than with potash feldspar, although Callaghan and Lemmon report orthoclase (not verified by the writer) as associated with the ore minerals. Because of the similarities in composition of the late magmatic and the ore solutions and of the sequential relations of the minerals formed in the granite during the post-consolidation stage and in the veins during the ore-forming stages, the conclusion seems reasonable that the late magmatic emanations may actually initiate the general cycle of mineralization and, therefore, may have an essential part in the mineralizing process.

The ore deposition apparently occurred in three successive stages, the first of which introduced base-metal, the second tungsten, and the third molybdenum. In addition to the base metals, the first of the ore-bearing solutions also carried large amounts of silica and some potash, antimony, sulphur, silver, and manganese. The succession of mineral deposition was quartz, sericite, pyrite, sphalerite, tetrahedrite, chalcopyrite, galena, and a little rhodochrosite. It is not certain that fluorine was contained in these solutions, but at least, it had become concentrated in the solutions remaining at depth so that fluorite was deposited in some abundance during the following tungsten stage.

The second stage of deposition was initiated only after the filling of quartz and base-metal sulphides had been rather extensively fractured. The reopened ore channels then permitted the movement of solutions containing some potash, considerable fluorine, tungsten, and manganese, and much silica. The minerals deposited included fluorite, sericite, quartz, and huebnerite, and small amounts of siderite, rhodochrosite, scheelite, and calcite.

During the third stage, which followed a less extensive reopening of the ore channels, the solutions, again containing potash, silica, and fluorine, deposited fluorite, considerable amounts

of muscovite and sericite, some pyrite and molybdenite, and scant quantities of chalcopyrite in veins within and close to the granite.[12]

Thus, potash, silica, fluorine, sulphur, and variable amounts and kinds of metals were contained in the late magmatic as well as in the later metallizing solutions. The presence of these substances in each surge of the solutions suggests a close genetic relationship between them. The mineral sequence, therefore, may bear out McKinstry's assertion that the alteration of the rock by magmatic emanations may be much like an early intense type of wall-rock alteration, the ores appearing during a later, cooler stage of the same process.

The Coeur d'Alene district offers a somewhat analogous situation. The monzonite-syenite bodies of that locality are regarded as the "ore bringers." However, it is not generally known that these intrusive bodies, originally of dioritic or noritic character, have been profoundly altered by magmatic emanations particularly rich in potash, and as a consequence the rock has been changed to monzonite and syenite. Locally, where the emanations passed out into the bordering siliceous Belt rock, they, too, have been changed to syenite.[13] Along fault zones at some distance from the intrusives, the country rock has been extensively sericitized,[14] apparently by potash-rich emanations that moved outward and upward into the colder rocks, where sericite rather than potash feldspar was the stable alteration mineral. Ore deposition came later and deposits were formed within and at the margins of the main intrusive body [15] as well as along faults cutting the zones of bleached, sericitized sedimentary rocks.[16]

[12] A detailed description of the minerals and their paragenesis as well as a comprehensive discussion of the genesis of the deposits are in the course of preparation.

[13] A paper dealing with the alteration of these intrusives is in the course of preparation.

[14] Shenon, P. J., and McConnel, R. H.: The silver belt of the Coeur d' Alene district, Idaho. Idaho Bur. Mines and Geol. Pam. 50: 6, 1939.

[15] Umpleby, J. B., and Jones, E. L., Jr.: Geology and ore deposits of Shoshone County, Idaho. U. S. Geol. Surv., Bull. 732: 32–38, 1923.

[16] Shenon, P. J., and McConnel, R. H.: op. cit., pp. 8–9; Umpleby, J. B., and Jones, E. L., Jr.: op. cit., pp. 41–115.

The emanations associated with the igneous intrusion must have had their source in magma below and the later ore-bearing solutions also must have had their source in the deeper magma body. The alteration produced by the late magmatic emanations may, as in the Blue Wing district, represent the earliest stage of the local mineralizing process.

In summary, the great batholiths that the writer has found to contain much associated "metamorphic" granite are barren of ore, but smaller bodies of intrusive rock that are magmatic, except for minor, late-stage metamorphic alteration, as in the Blue Wing and Coeur d' Alene districts, are associated with ores. These facts suggest that large-scale granitization in Idaho was not accompanied by ore formation, but that locally granitizing solutions, like ore-forming solutions, followed channels such as those found near magmatic granite bodies. The source of the granitizing solutions and the ore-bearing solutions is not the adjacent intrusive body but is deeper, the ore-bearing solutions, though closely associated, being distinctly later than the end-stage solutions of the granite itself. The absence of orebodies about the batholithic masses that have been cited seems to indicate that emanations even from deep sources, though capable of transforming rock, may be incapable of forming ore.

The relations of granite and ore in Idaho seem to support Loughlin's contentions that deposits associated with granitic rocks intruded during or soon after the closing stages of a major compression, even though the amount of volatile matter associated with the intrusion may be great, have but trivial quantities of sulphides; and that large metalliferous deposits are products of deep-seated differentiation subsequent to the intrusion of batholiths or stocks and are associated with igneous rocks intruded during the long period of uplift following that of major compression.[17] In Idaho the batholithic invasions as a whole took place at the close of the late Jurassic orogeny, although the Cassia batholith may have been intruded later; in contrast, the relatively

[17] Loughlin, G. F.: Comments on the origin and major structural control of igneous rocks and related mineral deposits. ECON. GEOL., 36: 683–687, 1941.

small masses of granite in the Blue Wing and of monzonite and syenite in the Coeur d' Alene district were intruded along local zones of structural weakness during early Tertiary time, and, therefore, long after the period of major compression. The writer expects to present more data on the origin and major structural control of the igneous rocks and the related ore deposits of Idaho at a later date.

ALFRED L. ANDERSON.

CORNELL UNIVERSITY,
ITHACA, NEW YORK,
May 13, 1942.

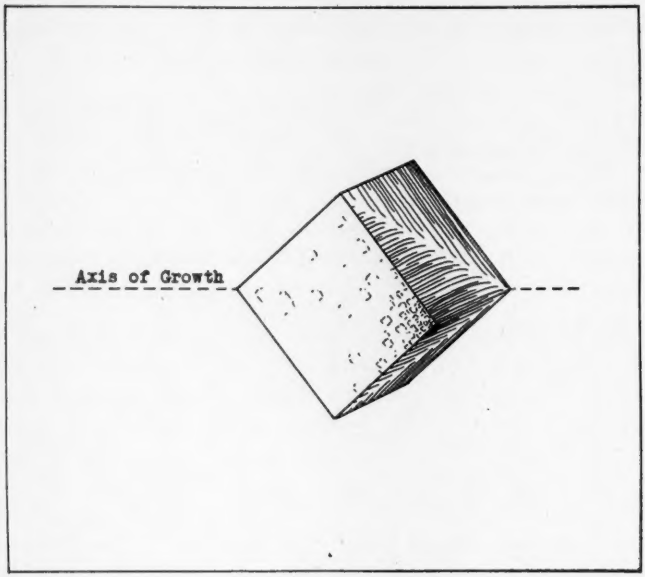
LINEAGE STRUCTURE AND CONDITIONS OF DEPOSITION OF PYRITE.

Sir: In recent papers on the variation of properties of pyrite [1] it was suggested that by combining the property variations with the manner of occurrence, a relation might be found that would permit a determination of the conditions of deposition of any given sample of pyrite. The properties studied were: electrical resistivity, temperature coefficient of electrical resistance, thermoelectric potential, composition, density, optical behavior, secondary crystal structure, crystal habit, color and luster. The observed variations were postulated to be due to small deviations from the ideal empirical formula, FeS_2, and also to the degree of development of lineage structure, the latter being of much greater importance in influencing the electrical properties.

The lineage structure observed in pyrite was related to the crystal structure as follows: one body diagonal of the cube was the axis of growth, and intersecting in this axis were three equivalent planes of growth that divided the crystal into three main segments; all the minor irregular prisms of growth in each segment were roughly parallel and were perpendicular to one of the cube faces. The accompanying diagrammatic sketch shows

[1] Smith, F. G.: Variation in the electrical conductivity of pyrite. Univ. Tor. Studies, Geol. Ser., 44: 83–93, 1940. Variation in the properties of pyrite. Amer. Min., 27: 1–19, 1942.

the observed relations. In general it was found that the larger the crystal of pyrite, the greater the development of lineage structure.



It was reported that as the degree of development of lineage structure of pyrite increased, the electrical resistivity increased, the temperature coefficient of resistance changed from positive, through zero, to negative, and the thermo-electric potential to copper changed from negative, through zero, to positive. While all three electrical measurements are very sensitive to changes in the lineage structure, only the thermo-electric potential can readily be measured on crystal fragments or crystal faces, since in these cases the potential is not influenced by the internal resistance of the crystal being studied. In addition, the thermo-electric potential is very easily and rapidly measured.

An instrument for measuring the thermo-electric potential of pyrite against metal has been made in the Department of Geology,

University of Toronto. It consists of a null-point potentiometer with a variable range and sensitivity, which measures the potential developed between two electrodes of stainless steel, one electrically heated, when they are touched to the surface of a crystal fragment of pyrite. During a series of measurements, variation in the temperature difference between the two electrodes is compensated by changing the total potential available to the balance bridge so that the galvanometer reads zero when the balance bridge is set at a pre-determined point and the electrodes are touched to a control crystal of pyrite.

The relation between the thermo-electric potential developed between the hot and cold electrodes in contact with a crystal of pyrite, and the lineage structure was observed on a crystal described and illustrated in the second of the two papers mentioned above. It is a large cubical crystal whose cube and octahedral planes had been ground flat and then etched to accentuate the lineage structure. The potential was found to be negative along the three growth planes, but rapidly became zero and then positive a few millimeters on either side of them. The largest positive values were found farthest away from any two adjacent growth planes. Thus while the average potential was large and positive, there were restricted zones within the crystal where it was large and negative. Another crystal, whose potential was found to be in general moderately large and negative, had small zones running in straight lines along which the potential was distinctly greater, and still negative. Such zones were probably traces of the three growth planes described above. Thus it appears that within any one crystal, fairly large differences in thermo-electric potential can be expected, depending on the spatial relation to the elements of the lineage structure: the nearer to the central axis or the three divisional planes, the more negative is the potential, and vice versa.

Notwithstanding the observed variation of thermo-electric potential within single crystals of pyrite, specimens from different localities showed a wide range of potential. Generally, the crys-

tal faces had a much more positive potential than the interior, but within the crystal the average potential was easy to determine. About one hundred specimens of pyrite from various deposits in the world were studied, and it appeared that they could be divided into three types with thermo-electric potentials that were large and negative, small and either positive or negative, and large and positive.

In the papers mentioned previously, it was pointed out that there is an apparent relation between the thermo-electric potential of a specimen of pyrite and the temperature at which it crystallized. Several examples were given, the relation being that pyrite formed at high temperatures (associated minerals, say pyroxene or specular hematite) had more negative potentials than pyrite formed at low temperatures (one example: pyrite encrusting Joplin sphalerite). Since a relation between thermo-electric potential and lineage structure has been postulated, there should then be a relation between temperature of formation and lineage structure. Such a relation is difficult to prove, but, in general, minerals that can crystallize over a range of temperature are found in nature much more perfect when they have crystallized at higher than at lower temperatures. Arborescent forms and radiating or divergent crystal structure, expressions of extreme development of lineage structure, are much more common in minerals deposited at lower temperatures. The physical reason for this condition is probably related to the greater amount of thermal agitation of the constituent atoms and the greater solubility of the minerals in the ore-depositing solutions at elevated temperatures, both of which would contribute to the growth of crystals with a minimum number of secondary crystal discontinuities.

The conclusion was reached that the degree of development of lineage structure of any crystal of pyrite can be determined in a relative manner by observing its thermo-electric potential against metal, and that crystals from different localities, deposited under different conditions, have significantly different potentials. The variation of the degree of development of lineage structure

of pyrite is tentatively related to the temperature at which it was deposited: the higher the temperature, the more perfect the crystal, and vice versa. Accepting this as a working hypothesis, the higher the temperature of deposition of pyrite, the more negative is its thermo-electric potential to metal.

The applications of this relation, which has yet to be rigorously proved, are very numerous and some should be of economic importance. For example, the general qualitative temperature history of a deposit containing several generations of pyrite could be determined. Also, if one generation of pyrite in a deposit is auriferous and the others barren, the identification of the former could be determined in detail without having recourse to frequent assays, so that the depositional controls of the auriferous variety could be observed. If the method proves useful, it could be made more exact in that pyrite deposited at temperatures determined by another method might be used to calibrate the potential scale. At present, the only point known with any certainty corresponds to the temperature of approximately 100° C.: the thermo-electric potential of pyrite from Joplin encrusting brown sphalerite.[2]

Information on the instrument used to measure the thermo-electric potential of crystals of pyrite will be supplied on request. Specimens of pyrite from deposits where the mineralogical relations have been studied in detail and especially where a clue exists as to the temperature of deposition, will be received with gratitude. Arrangements may be made to study specific mining problems related to different varieties of pyrite in a complex ore.

F. GORDON SMITH.

DEPARTMENT OF GEOLOGY,
UNIVERSITY OF TORONTO,
TORONTO, CANADA,
May 9, 1942.

[2] Newhouse, W. H.: Temperature of formation of the Mississippi Valley lead-zinc deposits. ECON. GEOL., 28: 744–760, 1933.

DIFFERENTIAL DENSITY OF GROUND WATER IN ORE DEPOSITION.

Sir: It seems to me that Mr. Brown [1] presents an idea to which much thought ought to be given. Certainly, as he points out, his theory seems particularly applicable to enriched ore deposits occurring at depths below the water table in the Canadian Shield area, where, as far as we know, the water table has always been high and relief low.

I should like to ask that if this theory is effective, would it not mean that enriched deposits should extend vertically, or at some small angle from the vertical, greater distances in some places than others in the same deposit, along zones of fissuring or more permeable beds. Most of the secondary deposits which I can recall offhand were pictured to me as essentially stratified horizontally. If Brown's theory has been effective I should think the picture would be anything but that. A brief discussion meeting this objection might be quite helpful in gaining wider acceptance of his idea.

Although Brown restricts his theory to those deposits in which the concept of artesian head has hitherto been applied, a word or two about the type of deposit to which it does not apply might be helpful.

It seems to me that this idea could be checked experimently with ease. Acidic solutions could be introduced into the upper part of a container filled with saturated material. In the upper sandy (unreactive) material metal filings could be placed for the acidic solutions to work upon. At some lower level, precipitation would be induced by reaction with limey material. Samples could be drawn off and analyzed at intervals at various levels by inserting horizontal tubes into the material. Another more simple set-up might be made on a colorimetric basis in which the movement of dyes introduced in acidic waters at the top of the container could be observed.

D. J. CEDERSTROM.

U. S. GEOLOGICAL SURVEY,
WASHINGTON, D. C.

[1] Brown, J. S.: Differential density of ground water as a factor in circulation, oxidation and ore deposition. ECON. GEOL. 37: 310–317, 1942.

REVIEWS*

Aerophotography and Aerosurveying. By LT. COL. JAMES W. BAGLEY, U. S. A. (Ret.). Pp. x + 324; Figs. 137. McGraw-Hill, New York, 1941. Price, $3.50.

This is the most up to date and complete work of its kind and merits the attention of anyone whose work is based on modern cartography. It "is intended for students who desire to make a systematic study of aerophotography and aerosurveying" and "is the result of studies carried out through many years and an arrangement of the substance of lectures given at the Institute of Geographical Exploration of Harvard University during the last four years."

The book opens with a discussion of some of the chief difficulties involved in aerial mapping together with a working terminology. It goes on to present material, with good illustrations and figures and ample mathematical analyses, on: aerial cameras and accessory instruments; lenses, color filters and shutters for aerial cameras; flights for photography—requirements and preparation; laboratory equipment and its use; reading and interpretation of aerial photographs; geometrical characteristics of aerial photographs; principles of photographic transformations and transforming printers; principles of stereoscopy and application to photogrammetry; methods of determining the tilt of an aerial photograph; methods of obtaining elevations and contours from aerial photographs; plotting methods and plotting instruments; mosaic maps and aerial photographs as illustrations; methods of obtaining horizontal and vertical control by ground surveys; map projections, grids and forms; finishing a map for reproduction; errors and measures for counteraction; calibration of cameras.

This bids fair to be "the Bible" of all who work with aerial maps, either in their preparation or in using them as a base for geologic mapping, and will undoubtedly be "the last word" on the subject for the duration. Improvements in technique by the military services will not be generally available until the present assignment is cleaned up.

BERNARR BATES, Lt., U. S. N. R.

* Books noted under *Reviews* and *Books Received* may be ordered through the Economic Geology Bookshop, W. S. Bayley, Urbana, Ill., but orders for official reports and single copies of Journals should be sent directly to their publishers.

Geomorphology: An Introduction to the Study of Land Forms, Pt. I. By C. A. COTTON. 3d Ed. Pp. 261; Figs. 254. Whitcombe and Tombs, Ltd., Christchurch, New Zealand, 1942. Price, 10 s 6 d.

A revised edition of the "Geology of New Zealand, Pt. I: Systematic," which presumably will be completed in Part II a little later, as the index references to items on pages number as high as 488. It is a well and clearly written elementary text based on the principles advocated by W. M. Davis, since "explanation is assumed to be a necessary part of landscape description." It is particularly interesting in that its many illustrations are nearly all of phenomena exhibited in New Zealand. The complete volume will contain about 505 pages and 474 illustrations, and will be issued in cloth binding at a cost of 25 shillings (about $4.25).

W. S. BAYLEY.

Weather Elements. By THOMAS A. BLAIR. Pp. 401; Figs. 107. Prentice-Hall, Inc., New York, 1942. Price, $5.00.

Here is an introduction to the science of meteorology presented concisely and systematically as a textbook for use by classes or by individuals whose daily work depends upon the weather. The language is far from being too technical for the layman and a reading should improve his understanding of that favorite topic of conversation. Each chapter is well illustrated and presents a list of problems as additional incentive for careful perusal. The topics include: The Atmosphere; Observing Temperature, Pressure, and Wind; Observing Moisture, Sunshine, Visibility, and Upper Air Conditions; Solar Radiation and Its Effects; Condensation of Water in the Atmosphere; Interrelations of Temperature, Pressure, and Wind; The General Circulation; The Secondary Circulation; Lesser Disturbances; Weather Forecasting; World Weather; Climate; Climate and Man; Electrical and Optical Phenomena; The United States Weather Bureau; and an Appendix with bibliography, conversion factors and tables, mean monthly and annual temperatures and precipitation tables, and a list of climatological section centers of the U. S. Weather Bureau.

It is strange that the author takes no cognizance of the work of Irving Krick and his associates in their highly successful long range forecasts for major airlines and the National Advisory Committee on Aeronautics, although a few pages are devoted to forecasting by analysis of air masses.

BERNARR BATES, Lt., U. S. N. R.

BOOKS RECEIVED.

WILLIAM E. BENSON.

U. S. Bureau of Mines, Washington, D. C.

Bull. 445. **Plastic and Swelling Properties of Bituminous Coking Coals.** R. E. BREWER. Pp. 260; Figs. 10. 1942. Price, 30 cents.

Bull. 446. **Typical Analyses of Coals of the United States.** A. C. FIELDNER, W. E. RICE, AND H. E. MORAN. Pp. 45. 1942. Price, 10 cents.

Bull. 447. **Quarry Accidents in the United States During 1940.** W. W. ADAMS AND V. E. WRENN. Pp. 87; Figs. 3. 1942. Price, 15 cents.

Bull. 448. **Coal-Mine Accidents in the United States: 1940.** W. W. ADAMS, L. E. GEYER, AND M. G. PARRY. Pp. 134. 1942. Price, 20 cents.

Techn. Pap. 626. **Analyses of West Virginia Coals.** A. C. FIELDNER ET AL. Pp. 341; Figs. 10. 1942. Price, 45 cents.

Techn. Pap. 629. **Collecting and Examining Subsurface Samples of Petroleum.** PETER GRANDONE AND A. B. COOK. Pp. 68; Figs. 24. 1941. Price, 20 cents.

Techn. Pap. 633. **Technical and Economic Study of Drying Lignite and Sub-bituminous Coal by the Fleissner Process.** L. C. HARRINGTON, V. F. PARRY, AND ARTHUR KOTH. Pp. 84; Figs. 32. 1942. Price, 20 cents.

Techn. Pap. 634. **Carbonizing Properties and Petrographic Composition of Lower Lignite-Bed Coal from the Atlas Mine, Middlesboro, Bell County, Ky., and the Effect of Blending This Coal with Pocahontas No. 3 and No. 4 Bed Coals.** J. D. DAVIS, D. A. REYNOLDS, G. C. SPRUNK, AND C. R. HOLMES. Pp. 47; Figs. 28. 1942. Price, 10 cents.

Techn. Pap. 637. **Index of Coals Tested in the Bureau of Mines Survey of Carbonizing Properties of American Coals.** J. E. WILSON AND J. D. DAVIS. Pp. 10. 1942. Price, 5 cents.

Techn. Pap. 638. **Photomicroscopy of Salt in Petroleum.** L. F. CHRISTIANSON AND J. W. HORNE. Pp. 56; Figs. 21. 1942. Price, 25 cents.

Techn. Pap. 639. **Bibliography of Bureau of Mines Investigations of Coal and Its Products, 1935 to 1940.** A. C. FIELDNER. Pp. 43. 1942. Price, 10 cents.

Techn. Pap. 640. **Coke-Oven Accidents in the United States During 1940.** W. W. ADAMS AND V. E. WRENN. Pp. 19; Figs. 3. 1941. Price, 5 cents.

Techn. Pap. 643. **Theoretical Calculations for Explosives. II. Explosion Pressures.** F. W. BROWN. Pp. 26; Figs. 12. 1942. Price, 10 cents.

Soil Survey of McKenzie County, North Dakota. M. J. EDWARDS AND J. K. ABLEITER. Pp. 99; 3 large soil maps in color; Figs. 3; Pls. 3. U. S. Dept. Agriculture, Washington, 1942. Price, $1.25.

Illinois Geological Survey. Reports of Investigations. Urbana, 1942.

No. 78. **Correlation of Domestic Stoker Combustion with Laboratory Tests and Types of Fuels. 1. Preliminary Studies.** L. C. MCCABE, S. KONZO, AND O. W. REES. Pp. 19; Figs. 9; Tables, 2.

No. 79. **Feldspar in Illinois Sands.** A Study of Resources. H. B. WILLMAN. Pp. 87; Figs. 13; Tables, 5.

No. 80. **Modern Concepts of Clay Materials.** RALPH E. GRIM. Pp. 50; Figs. 12.

No. 81. **Determination of Ash in Coals Unusually High in Calcite and Pyrite.** O. W. REES AND W. A. SELVIG. Pp. 10.

Canada Dept. Mines and Resources, Ottawa, 1942.

Publications of the Dominion Observatory. **Bibliography of Seismology No. 11, Jan. to June, 1942.** ERNEST A. HODGSON. Pp. 11.

Mines and Geology Branch, Bur. of Mines. No. 607. **Report on Investigations 1934–1940.** HARALD A. LEVERIN. Pp. 112; Figs. 2. Price, 25 cents.

Field Tests for the Common Metals. 8th Edit. GEORGE R. FANSETT. Pp. 56. Univ. of Ariz. Bull. Arizona Bur. Mines Techn. Series No. 40, Bull. 150. Tucson, 1942. Price, 25 cents.

Kansas Mineral Resources for Wartime Industries. JOHN M. JEWETT AND W. H. SCHOEWE. Pp. 180; Figs. 13; Tables, 26. Kansas Geol. Surv. Bull. 41, Lawrence, 1942.

Florida Geological Survey, Geological Bulletins, Tallahassee, 1942.

No. 20. **Stratigraphic and Paleontologic Studies of Wells in Florida, No. 2.** W. STORRS COLE. Pp. 89; Figs. 4; Pls. 16.

No. 21. **Geology of Holmes and Washington Counties, Florida.** ROBERT O. VERNON. Pp. 161; Figs. 20; Pls. 2; Tables, 8.

Brazil Div. de Fom. da Prod. Min., Rio de Janeiro.

Bol. 44. **Hygiene da Minas de Ouro Silicose Morro Velho-Minas Geraes.** C. M. TEIXEIRA, M. CURTY, E. MACEDO, AND O. BARBOSA. Pp. 158; Figs. 47. 1940.

Bol. 45. **Amianto No Brasil.** G. M. DE ARAUJO OLIVEIRA AND M. LISBÔA. Pp. 41; Figs. 6. 1940.

Bol. 47. **Salitre No Piauí.** G. M. DE ARAUJO OLIVEIRA. Pp. 91; Pls. 6. 1940.

Bol. 48. **Generalidades Sobre Cobre No Rio Grande Do Sul.** E. A. TEIXEIRA. Pp. 36; Pls. 3. 1941.

Bol. 49. **A Mina de Cobre de Camaquan Rio Grande Do Sul.** E. A. TEIXEIRA. Pp. 47; Pls. 4. 1941.

Magnetometer Surveys During 1941. E. L. TULLIS. Pp. 40; Figs. 9. South Dakota Geol. Surv. Rept. of Invest. No. 42. Mimeo. Vermillion, S. Dakota, 1942.

SOCIETY OF ECONOMIC GEOLOGISTS

ANNUAL MEETING, 1942.

The annual meeting of the Society will commemorate the centenary of the founding of the Geological Survey of Canada. It will be held on December 29th to 31st inclusive, at Ottawa, Canada, in association with the meeting of the Geological Society of America and affiliated organizations.

Members are urged to submit titles and abstracts of contemplated papers to one of the undersigned at their earliest convenience in order to secure space on the program. The closing date for consideration of such papers has been set as November 1st. Members are also earnestly requested to direct the attention of non-member geologists, engineers and geophysicists to this opportunity for presenting papers and for attending the meetings. Papers dealing with the mineral deposits of Canada, with the bearing of economic geology upon war-time production and with the military aspects of engineering geology will be of special interest. Forms for submitting abstracts may be obtained from the undersigned.

H. C. GUNNING,
Department of Geology,
University of British Columbia,
Vancouver, B. C., Canada,

M. E. WILSON,
Geological Survey of Canada,
Ottawa, Ont., Canada,

DAVID G. THOMPSON,
Water Resources Branch,
U. S. Geological Survey,
Washington, D. C.,

CHAS. H. BEHRE, JR.,
Department of Geology,
Columbia University,
New York, N. Y.

SCIENTIFIC NOTES AND NEWS

THE UNIVERSITY OF WEST VIRGINIA will dedicate its new million-dollar Mineral Industries Building at Morgantown on October 16th and 17th. The new building houses the school of mines, the West Virginia Geological Survey, the department of geology and the departments of chemical, metallurgical and ceramic engineering.

W. B. HEROY has been appointed director of reserves of the Office of Petroleum Coordinator. He takes over the duties of E. DE GOLYER who was recently appointed assistant deputy petroleum coordinator.

R. M. DREYER, assistant professor of geology and mining engineering at the University of Kansas, is on leave for active duty as Ensign in the U. S. Naval Reserve.

DAVID GALLAGHER was married to Miss Mary Dole on August 8th in Houston, Texas.

O. BOWLES has been named chief of the non-metal economics division of the U. S. Bureau of Mines to replace P. M. TYLER who resigned from the Bureau to join the Board of Economic Warfare.

J. J. COLLINS is now in the Spokane, Washington, office of the U. S. Geological Survey for investigation of strategic mineral deposits in the northwestern states "for the duration."

J. G. BARRY is head of a permanent field staff in Mexico for the metals and minerals division of the Board of Economic Warfare.

C. E. STOTT, formerly assistant general manager, has been promoted to vice president and general manager of the Cia. Minera de Peñoles and also Cia. Metallurgica Peñoles.

R. E. BROWN who has been junior geologist with the Oregon State Department of Geology and Mineral Industries is now junior geologist with the U. S. Geological Survey.

E. S. HILLS, associate professor of geology at the University of Melbourne, has been awarded the Wollaston Fund by the Geological Society of London.

R. S. MOEHLMAN of the Anaconda geological staff has established an office in Reno to supervise prospect drilling on Nevada copper properties.